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Abstract

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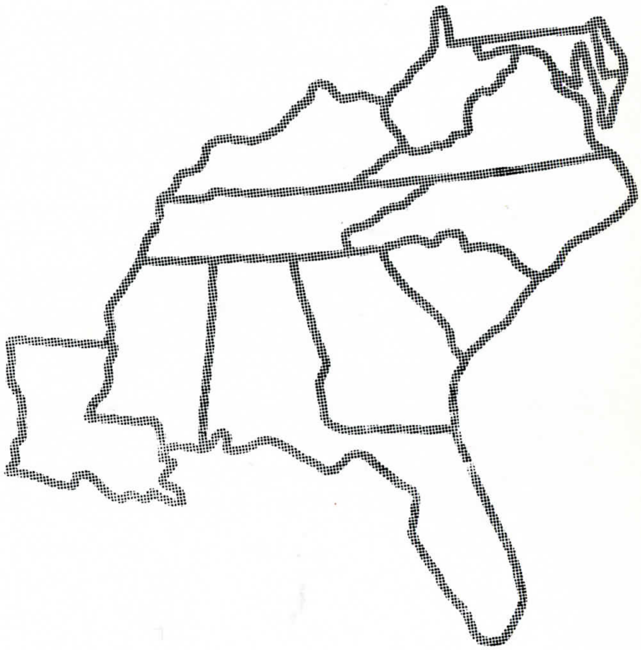

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STRATIGRAPHY OF THE NORTHERN CHARLOTTE METAMORPHIC BELT: APPLICATION TO THE CHARLOTTE BELT/MILTON BELT PROBLEM

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ABSTRACT

In 1955, the Charlotte belt was defined as the area of primarily plutonic rocks around Charlotte, North Carolina. Subsequent work has tended to ignore the pluton-based definition, and the Charlotte belt has been extended southwest from the type area into South Carolina and Georgia, and northeast into Virginia. These areas consist mainly of metavolcanic rocks with some metasedimentary rocks and subordinate intrusive rocks, and have been distinguished from adjacent belts by the amphibolite facies metamorphic grade, rather than the content of plutons.

In the 1980s, some workers returned to the plutonic basis of defining the Charlotte belt and, accordingly, gave a different name, the *Milton* belt, to the amphibolite facies metavolcanic rocks in northern North Carolina and Virginia northeast of the type area. Others have referred to this part of the belt as the Milton "terrane," and still others have proposed that the belt stretching from Georgia into Virginia, which most workers have traditionally called the Charlotte belt, be instead called the Charlotte *metamorphic* belt, and that the Charlotte belt and Milton belt/terrane be considered as components. The "Charlotte belt + Milton belt = Charlotte metamorphic belt" terminology is used in this paper.

Geologic mapping for this study in the south-central Virginia Charlotte metamorphic belt (Milton belt) indicates that the lithologies are mainly mixed felsic and mafic metavolcanic rocks, with subordinate metapelite and metagraywacke, and are part of the upright limb of a large recumbent nappe. The core of the nappe crops out in the vicinity of the Virginia/North Carolina border, and is somewhat more mafic than the overlying rocks to the northeast and southwest; additionally, the core sequence contains rare calc-silicate, marble, and quartzite. The country rock of the Charlotte belt type area is generally of the same type as the rocks to the north in the south-central Virginia portion of the belt. The metavolcanic lithologies and the amphibolite metamorphic grade are consistent characteristics along the entire length of the Charlotte metamorphic belt from Georgia through Virginia.

By contrast, the plutons of the Charlotte belt type area have been shown by other workers to represent several different magmatic episodes, range from late Precambrian to late Paleozoic, and to intrude most other southern Appalachian belts as well. The plutons therefore are not a suitable criterion by which to define the Charlotte belt or to distinguish it from other southern Appalachian belts. The consistency of host rock lithologic characteristics throughout the Charlotte belt/Milton belt leads to the conclusion that, regardless of the semantics, the belt is

part of a single larger lithotectonic unit defined by previous workers as the Carolina terrane.

INTRODUCTION

Purpose Of The Paper

The name Charlotte belt was first used by King (1955) for the abundantly plutonic region around Charlotte, North Carolina, in contrast with the primarily metavolcanic Carolina slate belt adjacent to the southeast (Figure 1). The northwestern boundary is more variable. In Virginia, the belt is separated from the eastern Blue Ridge belt by the Danville Triassic Basin. To the southwest, into North Carolina, South Carolina, and Georgia, the northwestern boundary is with the Inner Piedmont belt, the Kings Mountain belt, and again with the Inner Piedmont belt. Because these plutons are hosted by a belt of high-grade gneisses that border the low-grade Carolina slate belt and extend both northeastward into Virginia and southwestward into Georgia and Alabama, many workers after King applied the name "Charlotte belt" to the entire high-grade area (e.g., Tobisch and Glover, 1969, 1971; Hatcher, 1972; Williams, 1978). In an unpublished report, Butler (1980) named the northeastern extension of the gneissic host rocks of the Charlotte belt the "Milton belt," but did not propose another name for the southwestern extension. The primary basis of this distinction was the non-plutonic character of this area as opposed to that around Charlotte, North Carolina. Butler (1980) accordingly placed the Charlotte belt/Milton belt boundary at the contact between the plutonic suite and the predominantly gneiss and schist terrane to the northeast. Butler (personal communication, 1989) was also influenced by the apparent absence of any rock like granite of the 463±14 Ma (U-Pb, zircon, Hund, 1987) Shelton Formation (Henika, 1980; Kish, 1983) anywhere in the Charlotte belt southwest of the Milton belt.

Horton and others (1989) renamed the Milton belt the Milton "terrane," considering it to be a separate terrane from the Charlotte belt to the southwest. Recognizing that this view is not universal and that good precedent exists for calling the entire belt the Charlotte belt, Horton and Zullo (1991) proposed the name Charlotte *metamorphic* belt to encompass both the Charlotte and Milton belt portions of the belt (Figure 1). The name Charlotte metamorphic belt emphasizes the grade differential that has traditionally been used to distinguish the belt from the adjacent Carolina slate belt, while at the same time playing down the plutonic distinction that has become less useful as the variously aged plutons have now been better characterized. Because the names Charlotte belt and, more recently, the Milton belt have become common terms in the literature, it appears that acceptance of the Charlotte belt as the southern part of the Charlotte metamorphic belt and the Milton belt as the northern part may be useful in a geographic sense if no geologic connotation for the distinction is assumed. The purpose of this paper is to show that there are no significant differences between the Charlotte and Milton belts and that both, therefore, are part of the same terrane, the Carolina terrane (Secor and others, 1983).

Historical Development Of Research In The Charlotte Metamorphic Belt

In south central Virginia, Laney (1917) produced a map that included the

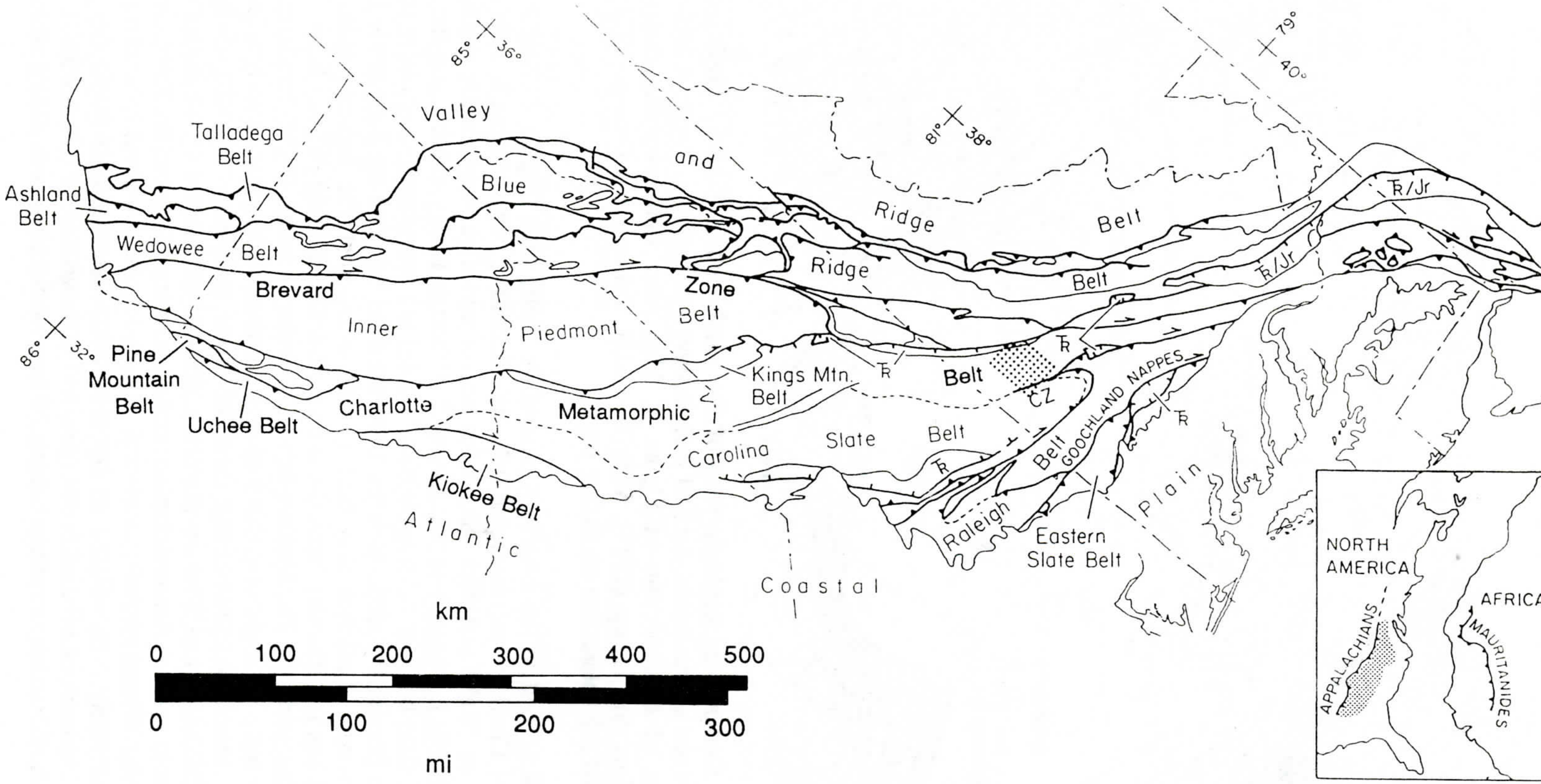


Figure 1. Map of the central and southern Appalachian orogen showing locations of major geologic belts. Triassic and Jurassic outcrops are marked by Tr and Jr. Shaded area in inset gives location with respect to North America. Dotted enclosure is study area in Figure 2. Other names are self-explanatory. Map from Glover (1989).

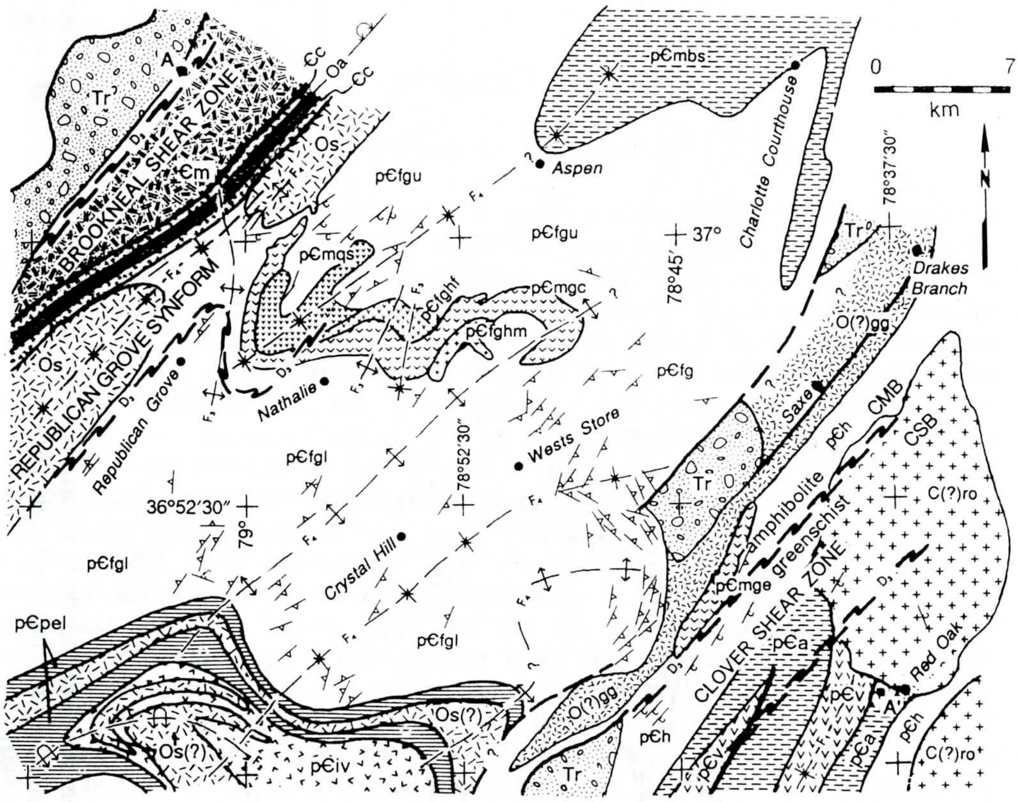


Figure 2. Generalized geologic map of the south-central Virginia Piedmont. All boundaries are approximate. Area of map is shown in Figure 1. Points A and A' are end points of cross-section in Figure 3. Map after field mapping during this investigation, and Laney (1917), VDMR (1963), Tobisch and Glover (1971), and Gates (1981). Explanation on facing page.

Carolina slate belt in the southeast part of the study area (Figure 2), and a small segment of Charlotte metamorphic belt. His Charlotte metamorphic belt rocks are described as "light and dark gray mica gneiss and dark amphibolite gneiss, the latter probably intrusive into the former." Although Laney did not use the terms "Carolina slate belt" or "Charlotte belt," he did use metamorphic grade to make a distinction between what is now the Carolina slate belt and the high-grade belt to the northwest. LeGrand and Bell (1966) noted that the metamorphic grade difference has traditionally been the means of differentiating the two belts.

In a report on the kyanite belt of Virginia, Jonas (1932) constructed a 1:125,000 scale geologic map that included portions of the present study area north of 37° N. Her work placed large portions of the metavolcanic rocks of the area into the Columbia and Shelton granites. She mapped much of the higher grade metavolcanic rock as metaepiclastic/metasedimentary Wissahickon Formation and extended the Wissahickon Formation as far southwest as the northwestern part of the present study area. Based on her work, the state geologic map of Virginia (Virginia Division of Mineral Resources, 1963) shows a large 130 x 8 km body of Shelton Granite extending from about 60 km northeast of the study area south-

EXPLANATION

STRATIGRAPHY

Central Charlotte Metamorphic Belt

Triassic
- Tr — *Arkose and conglomerate*

Ordovician
- Oa — *Arvonia Formation*
- Os — *Shelton Formation*

Cambrian
- Єm — *Melrose granite*
- Єc — *Chopawamsic Formation*

Precambrian
- pЄmbs — *Muscovite biotite schist unit*
- pЄfgu — *Upper felsic gneiss unit*
- pЄmqs — *Muscovite quartz schist unit*
- pЄfghf — *Fine-grained, hornblende-containing felsic gneiss unit*
- pЄfghm — *Medium-grained, hornblende-containing felsic gneiss unit*
- pЄmgc — *Mafic gneiss unit - central Charlotte metamorphic belt*
- pЄfgl — *Lower felsic gneiss unit*
- pЄpel — *Pelitic schists*
- pЄiv — *Intermediate volcanic rocks*

Eastern Charlotte Metamorphic Belt /Carolina Slate Belt

Triassic
- Tr — *Arkose and conglomerate*

Ordov.
- O(?)gg — *Granitic gneiss unit*

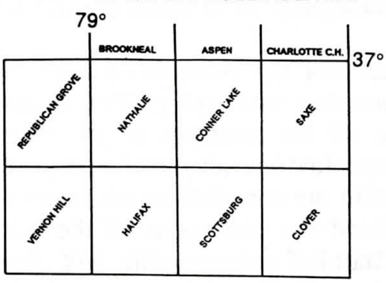
Cambr.
- Є(?)ro — *Red Oak granite*

Precambrian
- pЄv — *Virgilina Formation*
- pЄa — *Aaron Formation*
- pЄh — *Hyco Formation*
- pCmge — *Mafic gneiss unit - eastern Charlotte metamorphic belt*

SYMBOLS

- S2 regional foliation - selected to show structural trends
- Shear bands
- Overturned antiform
- Antiform axis
- Synform axis
- 1:24,000 topographic quadrangle map corner
- Boundary of rocks affected by shearing event
- Brittle fault
- Lithologic boundary
- Unconformity

TOPOGRAPHIC QUADRANGLE INDEX MAP



west to the North Carolina border. Work by Henika (1977) has subsequently reduced the area mapped as granite.

Brown's (1969) mapping in the Dillwyn, Virginia area, about 60 km to the northeast, included part of the Wissahickon near the northern tip of the Charlotte metamorphic belt. Slightly over half of his map area is of gneisses on strike with King's (1955) Charlotte belt type area. Brown referred to these gneisses as the "Hatcher Complex." The Hatcher Complex consists of metavolcanic and metasedimentary rocks that include biotite gneiss, hornblende gneiss, mica schist, and micaceous quartzites. Felsic gneisses, interlayered with hornblende gneisses and migmatite, are also present and are intruded to the east and northeast by the Columbia granite.

According to Conley (1978), the Hatcher complex includes part of the Chopawamsic Formation (Southwick and others, 1971), a metavolcanic/metasedimentary unit also composed of biotite gneiss, hornblende gneiss, mica schist, and micaceous quartzites. From the type area along Chopawamsic Creek in northern Virginia (Southwick and others, 1971), the Chopawamsic Formation has been progressively extended southwestward into north-central Virginia (Higgins and others, 1973; Pavlides and others, 1974, Conley and Johnson, 1975), and into central Virginia (Conley, 1978, Conley and Marr, 1979, 1980; Marr, 1980a, 1980b, 1981; Gates, 1981). The Chopawamsic Formation is time equivalent (550 Ma, U-Pb, zircon, Tilton and others, 1970; Higgins and others, 1971) with, and according to Glover (1989) is equivalent to, the younger Carolina slate belt sequence (Glover, 1974) in the central North Carolina Carolina slate belt. The younger slate belt sequence consists of the Uwharrie-Tillery-Cid-Floyd Church-Yadkin Formation sequence of metavolcanic/ metaepiclastic rocks of late Precambrian to Early Cambrian age (Milton, 1984; Harris and Glover, 1985, 1988; Goldsmith and others, 1988).

Unconformably overlying the Hatcher Complex, according to Brown (1969), is the Middle to Late Ordovician Arvonia Formation (Tillman, 1970). The Arvonia Formation is contained within the core of a tight, northeast-trending syncline. Later work to the north by Pavlides (1976, 1981) also indicated an unconformable contact between the Quantico Slate (Arvonia Formation equivalent) and the underlying Chopawamsic Formation. North of the James River, the base of the Arvonia Formation is marked by a conglomeratic quartz-mica schist (Brown, 1969) containing blue quartz pebbles (Conley and Marr, 1980).

Adjacent to the Dillwyn area (Brown, 1969) are the Andersonville (Marr, 1980a) and Willis Mountain (Marr, 1980b) 7.5-minute quadrangles (see also Marr, 1981). Marr's work has shown that most of the lithologies in these quadrangles are metasedimentary and metavolcanic in origin. The structurally lowest unit is a monotonous sequence of phyllites that may be in thrust contact with the overlying Chopawamsic Formation.

Marr's (1980a, 1980b, 1981) work showed that the Arvonia syncline, the schistose Arvonia Formation, and the unconformably underlying Chopawamsic Formation extended at least as far southwest as 40 km northeast of the present study area. Gates (1981) extended the Arvonia syncline into the northwest part of the study area, and mapped a graphitic schist within the core that he considered to be the Arvonia Formation. Glover (1989) pointed out that rocks of the Kings Mountain belt on the northwest side of the Charlotte metamorphic belt shares similarities with the Chopawamsic Formation/Arvonia Formation sequence and proposed that the two may be partially equivalent.

The Charlotte metamorphic belt has been mapped across nearly the entire width near the Virginia/North Carolina border by Henika (1977, western portion), Tobisch and Glover (1969, 1971, central and eastern portion), and Kreisa (1980, eastern portion). Tobisch and Glover (1969; 1971) concluded that during deformation Carolina slate belt and upper level Charlotte metamorphic belt rocks were folded and transported northwestward over lower level Charlotte metamorphic belt rocks forming a regional scale recumbent fold nappe (Figure 3). Conley (1985) referred to the Virginia Charlotte metamorphic belt as the south-

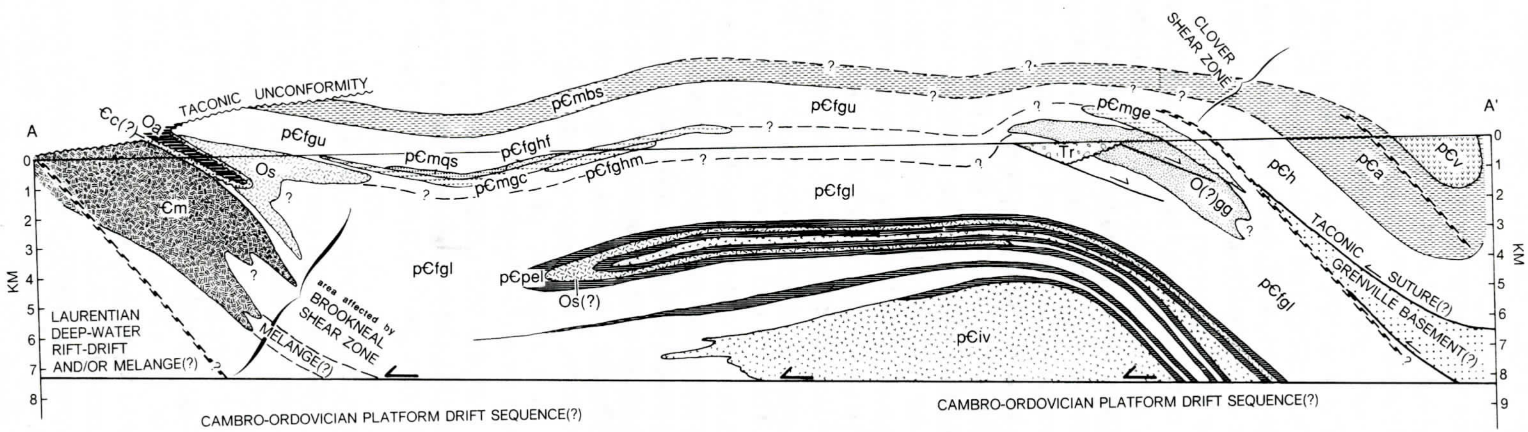


Figure 3. Cross section showing large recumbent fold nappe in the area of Figure 2. Above-surface geology is projected in from the northeast and deep structure is projected in from the southwest (Figure 2). Presence of platform sequence below thrust is based on interpretation of Interstate 64 seismic line to the northeast (Pratt and others,1988; Glover, 1989). Slice of Grenville(?) is included to take into account the Grenville Goochland nappe outcrop just northeast of the study area (Figure 1). The southeastern margin of the Brookneal shear zone is not shown so that thin formations in the area may be more readily apparent. Vertical and horizontal scales are the same.

central Virginia metavolcanic-plutonic belt, and noted that the rocks there were on strike with rocks called Charlotte metamorphic belt to the southwest, and accordingly believed that "these rocks probably are the country rock that the igneous plutons intruded to produce this igneous belt." Tobisch (1972) noted the predominance of large volumes of felsic gneiss and postulated that the felsic units represented highly metamorphosed felsic pyroclastic and volcaniclastic rocks and that some of the mafic gneisses represented mafic volcanic counterparts. Tobisch and Glover (1971) left open the possibility that Grenville basement could be present in the core of the nappe. The work of Henika (1977) basically agreed with the nappe interpretation of Tobisch and Glover (1971), whereas the work of Kreisa (1980) was generally consistent with the Tobisch and Glover (1971) stratigraphy to the east.

To the west of the Tobisch and Glover (1971) area, Henika (1977) divided the Charlotte metamorphic belt rocks into just two broad units. The low-grade of metamorphism (greenschist facies) he recognized near the Dan River Triassic basin enabled Henika to confirm the volcanic and epiclastic origin of the rocks. The stratigraphy he described consists of a lower unit of alternating mafic and felsic metavolcanic rocks and lenticular psammitic and pelitic metasedimentary rocks, and an upper unit of predominantly felsic metavolcanic rocks containing ash flow metatuffs. Henika (1977) placed the boundary between the two at the uppermost prominent mafic layer of the lower unit. Dikes and sills of composition similar to the metatuffs are present in both the lower and upper units and possibly represent feeders for the extrusive units (Henika, 1977).

Kreisa (1980) remapped at a larger scale part of the Tobisch and Glover (1971) area near South Boston, Virginia. As did Henika (1977), he condensed the various units of Tobisch and Glover (1971) and Tobisch (1972) into two broad categories, gneiss and biotite gneiss. The biotite gneiss unit he described consists of biotite-rich gneiss and subordinate biotite schist. Five to ten percent of interlayered mica schist and mica-rich gneiss suggested an epiclastic sedimentary component. The gneiss unit includes hornblende-biotite gneiss, biotite gneiss, and hornblende amphibolite and gneiss, with some interlayered mica schist. The chemical composition, small-scale interlayering, and lithologic variability indicated origin as a sequence of volcanic rocks. Glover and others (1971) obtained a 740 Ma date (U-Pb, zircon) from the gneiss unit. Rare high-alumina schist suggested that the sedimentary component was small. Harris and Glover (1988) considered that the amphibolite facies metamorphic rocks of the Charlotte metamorphic belt were probably equivalent to lower grade rocks in the Carolina slate belt.

Farther to the southwest into the Charlotte metamorphic belt of North and South Carolina, local investigations have been carried out, but many have been aimed at petrographic studies and age dating of the various intrusive rocks of the Charlotte metamorphic belt. These include the work of Bell and Overstreet (1959) and Butler and Fullagar (1978) in North Carolina, and of Fullagar (1971) in North and South Carolina. Butler and Secor (1991) and Maher and others (1991) provide reviews of the Charlotte metamorphic belt and other belts in North and South Carolina, and also refer to recent theses an other studies done in this part of the Charlotte metamorphic belt.

Bell and Overstreet (1959) postulated that many of the dikes in the Charlotte metamorphic belt could have been feeders for Carolina slate belt volcanic rocks. Stuckey and Conrad (1958) made no mention of possible volcanogenic components in the North Carolina Charlotte metamorphic belt, but Bell (LeGrand and Bell, 1966) believed that strong compositional layering implied that some of the rocks were derived from sedimentary or volcaniclastic protoliths.

Stratigraphic and petrographic work on the Charlotte metamorphic belt in South Carolina includes that of McCauley (1961), work on the crystalline rocks by Overstreet and Bell (1965a, 1965b), and of Gilbert and others (1982). McCauley (1961) noted that the Charlotte metamorphic belt was more mafic than the Carolina slate belt. Overstreet and Bell (1965b), and later Misra and McSween (1984) reasoned that because the sequence of rocks in the Kings Mountain, Charlotte, and Carolina slate belts (Figure 1) shared so many common features, the stratigraphy probably continued across all three belts, possibly with local intervening unconformities. King (1955) also postulated that Carolina slate belt equivalents may be present in the Charlotte metamorphic belt.

Overstreet and Bell (1965b) divided the rocks of the South Carolina Charlotte metamorphic belt into two broad categories, granitoid gneiss and mica gneiss, both of which have been variably intruded by granite, pyroxenite, norite, syenite, and mafic dike swarms. The granitoid gneiss consists principally of feldspar biotite schist, granitoid gneiss and granitoid ranging in composition from granite to granodiorite, and feldspar biotite schist and gneiss with variable muscovite and hornblende. The mica gneiss is composed of biotite and hornblende gneisses and schists and associated granitoid rocks that closely resemble the rocks of the granitoid gneiss unit. Overstreet and Bell (1965a, 1965b) mapped a distinctive mafic unit at the base of the Carolina slate belt sequence that they extended into the Charlotte metamorphic belt as the boundary between the lower granitoid gneiss sequence and the upper gneiss unit. The large areal extent of the unit argues for formation as a flow or series of flows.

Secor and others (1986) noted that work in South Carolina suggests that the Charlotte metamorphic belt there is the infrastructure beneath a Carolina slate belt suprastructure. In contrast, Dallmeyer and others (1986) concluded that the Charlotte metamorphic belt in South Carolina contains a plutonic metaigneous complex that developed as a sub-volcanic infrastructure contemporaneously with the volcanic rocks in the overlying Carolina slate belt.

STRATIGRAPHY

The southeast portion of the Charlotte metamorphic belt of the study area has been intruded by what is now granitic gneiss (Figure 2). The full extent of this unit has not been mapped but, in the study area, it divides the main body of amphibolite facies Charlotte metamorphic belt rocks northwest of the granitic gneiss unit from a smaller body of such rocks southeast of it. The latter grades into greenschist facies Carolina slate belt rocks farther southeast (Figure 2). The granitic gneiss unit is used to divide the study area into two portions, the central Charlotte metamorphic belt area and the eastern Charlotte metamorphic belt/Carolina slate belt area (Figure 4). These portions are discussed separately below.

Central Charlotte Metamorphic Belt

The Charlotte metamorphic belt is lithologically variable and the units as described contain small proportions of other lithologies, e.g., local metamafic layers within a primarily metafelsite sequence. Each unit has been mapped as the

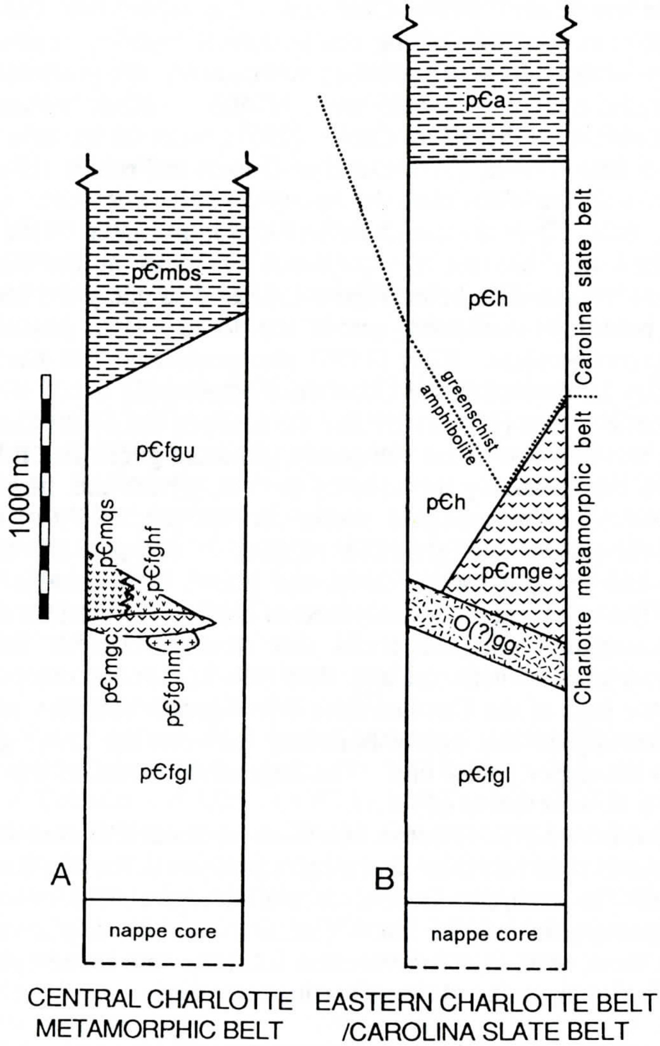


Figure 4. Stratigraphic columns showing representative cross-sections from two areas of the map in Figure 2. (a) Stratigraphic column of the northwestern 75 percent of the map area (Charlotte metamorphic belt). (b) Stratigraphic column of the southeastern 25 percent of the map area (eastern Charlotte metamorphic belt/Carolina slate belt). Unit boundaries are constructed to illustrate thickness ranges of the units. Transition from amphibolite facies rocks to greenschist facies rocks marks the Charlotte metamorphic belt/Carolina slate belt boundary. See text for descriptions of units.

dominant lithology over a given portion of the map area. Although basic rock types can be easily defined, the amphibolite facies metamorphic grade of metamorphism and consequent recrystallization does not permit identification of original volcano-sedimentary structures such as lapilli, flow banding, grading, and similar features.

Based on field mapping and modal mineralogy (Table 1), the principal rock type is biotite quartz plagioclase gneiss, including lesser amounts of K-feldspar

Table 1. Modal Analyses of Selected Representative Charlotte Belt Lithologies

SAMPLE	QTZ	PLAG	KSP	BIOT	CHLR	MUSC	HORN	EPID	SPHN	OPAQ	TOTAL[1]	COUNT
Lower felsic gneiss unit (K-feldspar plagioclase quartz gneiss)												
RB7- 5	34	39	20	2	--	3	--	1	--	--	99	247/600[3]
Mafic gneiss unit (central Charlotte metamorphic belt)												
RB6- 1	--	19	--	--	--	--	53	28	tr	--	100	600[2]
RB6- 25	10	--	--	--	--	--	42	47	1	--	100	600
RB6-423	19	--	--	--	6	--	26	47	2	--	100	600
Medium-grained hornblende-containing felsic gneiss unit(tonalite[4])												
RB6-182	31	54	--	7	--	--	1	5	--	1	101	700
RB6-183	20	52	--	11	--	--	12	2	--	3	100	600
RB6-191	30	55	--	--	9	--	--	6	--	--	100	600
Muscovite quartz schist unit												
RB6- 36	57	--	--	--	--	43	--	--	--	tr	100	600
RB6-270	74	--	--	--	--	26	--	--	--	--	100	600
RB6-380	66	--	--	--	--	34	--	--	--	--	100	600
Upper felsic gneiss unit (biotite quartz plagioclase gneiss)												
RB6- 14	38	38	--	10	--	--	3	12	--	--	101	600
RB6-139	46	39	--	8	--	2	--	5	--	--	100	600
RB6-140	33	54	--	11	--	2	--	1	--	--	101	600
RB6-175	37	55	tr	6	--	--	--	--	--	2	100	238/600
RB6-179	37	50	--	12	--	tr	--	tr	--	--	101	600
RB6-855	48	37	--	14	--	--	--	tr	--	1	99	600
Upper felsic gneiss unit (K-feldspar plagioclase quartz gneiss)												
RB6- 6	31	47	20	2	--	--	--	--	--	--	100	240/800
RB6-136	47	22	17	1	--	10	--	3	--	--	100	240/600
Muscovite biotite schist unit												
RB6- 9	35	29	--	21	--	14	--	--	--	2	101	600
RB6- 45	23	49	--	26	--	tr	--	--	--	2	100	600
Mafic layer in the muscovite biotite schist unit												
RB6- 53	20	7	--	--	--	--	--	69	--	4	100	600
Mafic gneiss unit (eastern Charlotte metamorphic belt)												
RB6-684	5	31	--	tr	--	--	62	2	--	--	100	600
RB6-729	9	51	--	2	--	--	35	2	--	1	100	600
Granitic gneiss unit (granite[4])												
RB6- 76	36	34	30	--	--	tr	--	--	--	--	100	700
Red Oak granite (granodiorite[4])												
RB6-100	29	49	22	--	--	--	--	--	--	tr	100	600

[1]Some analyses do not total 100 percent due to rounding

[2]Indicates total number of points counted on petrographic microscope

[3]First number is number of points counted by microprobe in order to distinguish fine-grained quartz, plagioclase, and K-feldspar. Second number is number of additional points counted on petrographic microscope to determine percentages of all phases other than quartz, plagioclase, and K-feldspar.

[4]Classification based on Streckeisen (1973)

plagioclase quartz gneiss. These rocks comprise slightly more than half of the stratigraphic column of the area. Metamafic units represent about a fifth of the column. Muscovite biotite schist makes up somewhat less than a fifth of the area, whereas muscovite quartz schist constitutes less than a tenth. The total structural thickness of these units in the study area is about 3.5 km. Because the rocks have been variably deformed, this thickness and others given below are considered as a present structural thickness, rather than an original stratigraphic thickness.

Mapping and structural analysis by Tobisch and Glover (1971) in the area near the Virginia/North Carolina border led them to conclude that the Charlotte metamorphic belt was a large recumbent fold nappe there. This interpretation was confirmed by the author (Baird and Glover, in review) in the adjacent area to the north based partially on the fact that the Charlotte metamorphic belt here is a large area (about 34 km across) characterized by a sub-horizontal foliation. The core of the nappe (Figure 3) is defined here as those rocks lying structurally below (southwest of) the lower felsic gneiss unit (Figure 2). According to Tobisch and Glover (1971), the dominant and oldest rock of the core is interlayered hornblende plagioclase gneiss and quartz feldspar gneiss with rare calc-silicate gneiss and marble. This unit is designated as intermediate metavolcanic rocks in Figure 2. The hornblende plagioclase gneiss unit of the nappe core yields an age of 740 Ma (U-Pb, zircon; Glover and others, 1971). Overlying this unit is a quartz feldspar gneiss unit with discontinuous layers of pelitic schist, quartzite, and less common calc-silicate gneiss (pelitic schist unit of Figure 2). These lithologies enclose medium- to coarse-grained granitic rock that is conformable within the hinges of folds related to the nappe. According to Kreisa (1980), the hornblende plagioclase gneisses of Tobisch and Glover (1971) include gneisses of both basaltic and andesitic composition; the more felsic rocks are of rhyodacitic composition. Farther to the southwest and into North Carolina, the lower felsic gneiss unit wraps around the core sequence.

The lower felsic gneiss unit (Table 1) of the central Charlotte metamorphic belt lies structurally above the metapelites and hornblende- and quartz-plagioclase gneisses of the nappe core (Figure 2), and below the mafic gneiss unit (Figure 4). It is primarily leucocratic, with a salt and pepper appearance caused by disseminated fine-grained biotite (0.125-0.25 mm). It is locally gradational with biotite-rich portions that Kreisa (1980) called biotite gneiss and interpreted as volcanic in origin. The biotite-rich gneiss has a chemical composition similar to that of graywacke (Kreisa, 1980) and is equivalent to the biotite-rich portions of the felsic gneiss unit. On a large scale, however, the unit is quite homogeneous. Total structural thickness is about 1100-1200 m.

Structurally above the lower felsic gneiss unit lies a mafic gneiss unit (Table 1) that extends across two 7-1/2-minute quadrangles in the area before eventually pinching out (Figure 2). The mafic gneiss is dark green to black, and fine-grained (0.125-0.25 mm). Locally, it contains concordant layers of felsic material and muscovite and biotite schist. The structural thickness of this unit is about 30-100 m, and averages around 70 m.

The fine-grained hornblende-containing felsic gneiss unit (Table 1; Figure 2) is a leucocratic, fine-grained (0.125-0.25 mm) gneiss. It has a salt and pepper texture like the lower felsic gneiss unit, but is primarily due in this rock to hornblende rather than biotite. The unit is about 25-30 m in structural thickness. The medium-grained hornblende-containing felsic gneiss (Table 1) is similar to the fine-grained unit, and lies about a kilometer to the southeast of it (Figure 2) . It is leucocratic to medium-gray, and medium-grained (0.5-2.0 mm). The outcrop pattern of the medium-grained unit is somewhat irregular, and appears to cross-cut the mafic gneiss unit. At least one exposure contains fine-grained mafic enclaves similar to the mafic gneiss.

The muscovite quartz schist unit (Table 1) has an irregular outcrop pattern (Figure 2). It is leucocratic and fine-grained (0.125-0.25 mm). Pyrite is common, and grains of chalcopyrite are locally associated with the pyrite. In thin-section, fine-grained garnet has been superimposed on the muscovite grains. These garnets are commonly color-zoned, with pinkish cores and clear rims. Locally, the schist contains grains of the zinc spinel gahnite. Average structural thickness of this unit is about 300 m.

The upper felsic gneiss unit (Table 1) is distinguished for convenience purposes only from the lower felsic gneiss unit on the basis of the partially intervening mafic gneiss unit (Figure 2). It is lithologically the same as the lower felsic gneiss unit. The structural thickness of the upper felsic gneiss unit is approximately 900-1000 m.

The muscovite biotite schist unit (Table 2) is medium to dark gray, and is generally coarse to locally very coarse-grained (0.5-2.0 mm), and commonly contains 0.5 cm or so megacrystic grains of feldspar in the matrix. The unit is up to 1100 m thick in the area, although the upper contact has not been mapped. The schist commonly contains layers of less biotite-rich gneiss. Part of this unit was included in the Wissahickon Formation by Jonas (1932).

Eastern Charlotte Metamorphic Belt/Carolina Slate Belt

The granitic gneiss unit (Table 1; Figure 2) is leucocratic, and fine-to medium-grained (0.5-2.0 mm). It is massive in places, but typically has a weak foliation. In the vicinity of the Triassic basins, the plagioclase is generally altered and exhibits a cataclastic texture.

Within the eastern Charlotte metamorphic belt Hyco Formation near the contact with the granitic gneiss unit (Figure 2) is a mafic gneiss unit with a structural thickness of about 500-1000 m, and averages about 725 m. It is dark green to black, and the dominant lithology is a fine- to medium-grained (0.125-0.50 mm) gneiss, with a few 1 mm or so feldspars in the matrix that give the rock a speckled appearance. The subordinate lithology is a fine-grained (0.125-0.25 mm) gneiss, with a salt and pepper texture and lower hornblende content very similar to the fine-grained, hornblende-containing gneiss of the central Charlotte metamorphic belt.

Other Units

Coarse-grained mafic gneiss bodies occur throughout the Charlotte metamorphic belt. These are generally too small to illustrate on the geologic map, but are somewhat more common in the northeast part of the area. They are coarse- to very coarse-grained (1.0-10.0 mm), and range from gabbroic in composition to hornblendite. The hornblendite is typically the coarser-grained rock. Foliation is generally weak. Locally, the coarse-grained mafic gneiss may be interlayered with fine-grained mafic gneiss.

Granitic to pegmatitic dikes, as well as small quartzofeldspathic pods and segregations are common throughout all the lithologies. They are, however, most common in more biotite-rich lithologies such as the upper felsic gneiss and muscovite biotite schist units. The mineralogy consists of microcline, somewhat subordinate plagioclase, and 30 percent or more quartz. The lithology is mica-poor, but where present is nearly always muscovite. Foliation is generally weak to non-existent. Grain size is coarse to pegmatitic, and commonly grades from one to the other in the same exposure. Locally, enclaves of foliated metamorphic rock are present.

Shelton Formation: The Shelton Formation, originally the Shelton *granite* (Jonas, 1932), consists, however, of both granitic and volcanic rocks. For this reason, Henika (1980) classed it as a formation. It is primarily a coarse-grained granite gneiss that is strongly sheared, exhibiting a classic L-tectonite fabric, with a strong subhorizontal mineral lineation (elongated grains of quartz and feldspar) and only a very weak southeast-dipping mylonitic foliation. It has been dated by Hund (1987) at 463±14 Ma (U-Pb, zircon) and by (Kish (1983) at 429 Ma (Rb-Sr, whole rock). As noted by Tobisch and Glover (1971) and Henika (1977, 1980), the Shelton Formation is exposed in the cores of regional synformal and antiformal folds.

Arvonia Formation: The Arvonia Formation is a silver-gray graphitic quartz muscovite schist of probable pelitic protolith. It has a paleontologically-determined age of Middle to Late Ordovician (Tillman, 1970), and unconformably overlies the younger Carolina slate belt-equivalent Chopawamsic Formation (Glover, 1989). Gates (Gates and others, 1986) extended the Arvonia Formation into the study area from about 40 km northeast where it had been mapped by Marr (1980a, 1980b, 1981) and is a schist. In this study, the schist has been mapped an additional 15 km to the southwest where it is inferred to be overlapped by sediments of the Danville Triassic basin. The Arvonia Formation is presently preserved only in a strip on the far northwestern edge of the map area. the structural thickness ranges to about 70 m.

DISCUSSION

Stratigraphy

Based on a number of factors, the original protoliths of most of the Charlotte metamorphic belt felsic and mafic gneisses examined are believed to be volcanic in origin. These factors include (1) the modal mineralogy (Table 1), (2) the fine-grained nature of the rocks, (3) the areally homogeneous character and (4) the great structural thickness. They are interbedded with mica schists that are probably pelitic/epiclastic in origin. An overall lack of cross-cutting relationships with other units appears to rule out a shallow intrusive origin. Supporting a volcanic origin interpretation is mapping by Henika (1977) in the Charlotte metamorphic belt to the southeast near the North Carolina border where the metamorphic grade is greenschist facies and the original volcanic textures of the rocks have not been obliterated by recrystallization. Specifically, the lower and upper felsic gneiss units, the fine-grained, hornblende-containing felsic gneiss unit, and the mafic gneiss units of the central and eastern Charlotte metamorphic belt are believed to be volcanic in origin.

The medium-grained, hornblende-containing felsic gneiss is considered to have been a tonalitic intrusive based on modal mineralogy (Table 1; Streckeisen, 1973), the relatively coarse grain size, and the presence of enclaves of dissimilar (fine-grained mafic gneiss) rock. The outcrop pattern showing that it cross-cuts the mafic gneiss unit of the central Charlotte metamorphic belt (Figure 2) is consistent with an intrusive origin. The mafic gneiss enclaves are interpreted to be xenoliths derived from the mafic gneiss unit. Based on the close proximity of the medium-grained, hornblende-containing felsic gneiss unit to the compositionally similar fine-grained, hornblende-containing felsic gneiss unit, the former may be the intrusive equivalent of the latter.

The muscovite quartz schist unit is virtually a bimineralic rock (quartz and muscovite; Table 1), contains unusual minor components such as sulfides and gahnite, and it exhibits an irregular outcrop pattern with respect to the adjacent stratigraphy (Figure 2). It is interpreted as a metamorphosed hydrothermally-altered felsic volcanic rock, such as would be found near a subaqueous volcanic vent where circulating hydrothermal waters have leached the feldspars and altered them to clay. The irregular outcrop pattern of this unit (Figure 2) is also consistent with the inferred hydrothermal origin. Because the schist is laterally continuous with the upper felsic gneiss unit, the former is probably equivalent to part of the latter. The gahnite may represent a concentration of zinc by the circulating waters; the sulfides probably have a similar origin. Based on the close proximity of the muscovite schist unit to the fine-grained, hornblende-containing gneiss unit, it is possible that alteration may be related to eruption of that unit.

Because of the micaceous nature of the muscovite biotite schist unit, the protolith is interpreted to have been a fine-grained sedimentary unit, representing epiclastic rocks and/or fine distal products from volcanic eruptions. The local mafic layers are interpreted as small sills or dikes, as their thinness and limited extent does not seem to be amenable to interpretation as a flow.

The granitic gneiss unit is interpreted as an intrusive because of the modal mineralogy (Table 1; Streckeisen, 1973), the grain size that is coarser than the surrounding fine-grained gneisses, and the massive to weakly foliated texture. The age and affinity of the granitic gneiss unit are unknown, but based on the weak foliation, it is probably a late syn-tectonic to post-tectonic intrusive.

Nature Of The Milton Belt

Much of the Milton belt in North Carolina and the southernmost part of the belt in Virginia consists of the core of a large recumbent fold nappe defined by Tobisch and Glover (1971). The core rocks are defined as such because they are structurally lowest in the area, they have been dated at 740 Ma (U-Pb, zircon; Glover and others, 1971), and the metamorphic grade is highest (amphibolite facies) compared with the greenschist facies Carolina slate belt to the southeast and local occurrences of greenschist facies rocks on the northwest side of the Charlotte metamorphic belt (Henika, 1977). Minor calc-silicate rocks are present in the biotite gneiss and schist unit in the Milton belt on the North Carolina geologic map (North Carolina Geological Survey, 1985). Such rocks are not present in the metafelsites and mica schists examined in the Milton belt of south-central Virginia. Tobisch and Glover (1969, 1971), however, described calc-silicates in the sequence of rocks they mapped in the nappe core which includes part of the sequence shown on the North Carolina state geologic map (North Carolina Geological Survey, 1985). The calc-silicates are only a minor component the nappe core, and the rocks of the nappe core lithologically are not significantly different from the rocks of the sequence overlying the core. With respect to the rock chemistry, the core rocks belong to the same calc-alkaline differentiation trend as do the rocks of the overlying sequence to the northeast (Baird, 1989; Baird, in review), which are volcanic rocks similar to the host rocks for the intrusive complex to the southwest around Charlotte, North Carolina.

Nearly four decades of research on the plutonic suite around Charlotte, North Carolina since the work of King (1955) has shown that the plutons are not as simple and homogeneous as he may have thought. Plutonism in the Charlotte belt, for example, ranges from late Precambrian through Permian (Sinha, 1989), and belongs to several different and unrelated tectonic events. Similar-age plutons are present in most other Piedmont belts of the southern Appalachian orogen and are by no means restricted to the Charlotte belt type area. The inappropriateness of using plutons to define the Charlotte belt is perhaps best exemplified by the area encompassed by the towns of High Point, Roxboro, and Burlington on the North Carolina state geologic map (North Carolina Geological Survey, 1985). This area is nearly 100 percent plutonic, is contiguous with King's (1955) type Charlotte belt area just to the southwest, and yet is included in the Carolina slate belt on the map.

The Milton belt is disqualified as an independent terrane because of the nature of the boundary with contiguous rocks. Only the northwestern boundary, which has been interpreted as the Taconic suture (Glover, 1989) can be classed as a fault separating fundamentally different rock bodies. The southeast boundary with the Carolina slate belt is a metamorphic isograd that has experienced a late Paleozoic dextral shear overprint (Baird, 1989; Baird and Glover, in review). Detailed work along some 11 km of this boundary indicates that the Hyco Formation occurs on both the Charlotte metamorphic belt and Carolina slate belt sides, indicating that the boundary is an overall conformable one. Although the boundary of the Charlotte metamorphic belt/Carolina slate belt trends into the the Scottsburg Triassic basin (southeast portion of Figure 2), there is no faulting directly associated with the boundary as Kreisa (1980) described southwest of the Scottsburg basin. The southwestern boundary of the Milton belt with the Charlotte belt type area is locally an intrusive, not a fault, contact. Where the host gneisses are still present, they are continuous from the Milton belt into the Charlotte belt. To the northeast, the belt continues on into northern Virginia where the lithologies have an uncertain relationship with the Chopawamsic volcanic sequence there.

Aside from a very few occurrences of fossils preserved in low-grade metamorphic rocks of the Carolina slate belt (e.g., Cloud and others, 1976; Secor and others, 1983), all terrane definition in the southern Appalachians must be done strictly based on isotopic ages, lithology, and the nature of contacts between rock bodies. Neither changes in lithology nor fault contacts exist within or bordering the Milton belt that would suggest that the rocks in which those lithologic changes or faults occur were once distinct geologic entities separated in time and/or space. On the basis of the present mapping, the Charlotte metamorphic belt of Virginia is best considered to be contiguous with and part of the Carolina terrane.

Given the present state of knowledge, it appears that the Charlotte metamorphic belt protoliths primarily consist of felsic volcanic rocks and epiclastic/volcaniclastic rocks, together with lesser mafic volcanic rocks, along the entire length of the belt. Because of this, the name Milton belt does not appear useful in a tectonic sense. However, because the name Milton belt has been incorporated into the literature and onto geologic maps of the area, it, along with the Charlotte belt, may serve as a geographic term, and the name Charlotte metamorphic belt, as proposed by Horton and Zullo (1991) may be applied along the entire length of the belt.

The Charlotte belt and Carolina slate belt are part of the Carolina terrane (Secor and others, 1983). The Milton belt is equivalent to the Charlotte belt and gradational with the Carolina slate belt here, thus the Milton belt is also part of the Carolina terrane as are the Charlotte belt and Carolina slate belt. Work in the southern Appalachians will eventually result in the elimination of many belt names, and their replacement with terrane names that are representative of the tectonic settings of the original lithotectonic entities.

CONCLUSIONS

1. A mappable stratigraphy exists in the Charlotte metamorphic belt of south-central Virginia. The stratigraphy consists of a metafelsite-metamafic-metafelsite-mica schist sequence that structurally overlies the rocks of the core of the large recumbent fold nappe mapped by Tobisch and Glover (1971). All rocks examined in this study are part of the upright limb of the nappe. The rock types are basically the same lithologies as the host metavolcanic rocks for the plutonism to the southwest in the Charlotte belt type area near Charlotte, North Carolina, and are gradational to the southeast with the Carolina slate belt.

2. Because there are no fundamental lithologic differences by which the Milton belt may be distinguished from the Charlotte belt host lithologies, and because the Milton belt shows little contrast with lithologies of adjacent belts along much of the "margin" and is not an entirely fault-bounded entity, it is recommended that the Milton belt be considered as equivalent to the Charlotte belt, and thus a part of the Carolina terrane. It is recommended that the name Charlotte metamorphic belt be applied to the entire high-grade belt northwest of the Carolina slate belt as delineated on maps such as those by Williams (1978), and Morgan (1982).

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IMPURE K-BENTONITE BEDS FROM THE LEXINGTON LIMESTONE AND THE POINT PLEASANT FORMATION (MIDDLE ORDOVICIAN) OF NORTHERN KENTUCKY AND SOUTHWESTERN OHIO

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ABSTRACT

Two stratigraphic zones containing impure K-bentonite beds are recognized within the Lexington Limestone and Point Pleasant Formation in the subsurface of northern Kentucky and southwestern Ohio. In the Lexington Limestone, two to four impure K-bentonite beds occur within 7.9 meters (26 feet) of strata herein named the Westboro K-bentonite zone. In the Point Pleasant Formation, two to three impure K-bentonite beds occur within 1.2 meters (4 feet) of strata herein named the Bear Creek K-bentonite zone.

Geophysical wireline logging, X-ray diffraction, petrological and light-and heavy-mineral analyses have documented the occurrence of these impure K-bentonite beds in nine cores from northern Kentucky and southwestern Ohio. Gamma ray, neutron, caliper, and bulk density geophysical logs show distinctive signatures which correspond to the stratigraphic position of impure K-bentonite beds in one of the two core holes logged. These beds contain ordered illite-smectite, untwinned feldspar, idiomorphic zircon, and apatite which are of volcanic origin. These minerals are mixed with calcite, quartz, dolomite, and collophane of detrital origin.

The impure K-bentonite beds occur as three discrete lithologies: 1) fossiliferous claystones, 2) biotite-rich carbonate and clastic rhythmites, and 3) burrow-fill lithologies. Storm sedimentation is interpreted as the physical mechanism producing lithologies 1 and 2. Lithology 3 is the result of burrowing organisms ingesting and redepositing the pre-existing mixture of volcanic ash and unconsolidated marine sediments.

The impure K-bentonite beds of the Westboro and Bear Creek K-bentonite zones occur in rocks assigned to the Middle Ordovician, Shermanian Stage of the Mohawkian Series. Exact position of these beds within the Shermanian Stage has not been established. Thus, lateral correlation of the Westboro and Bear Creek impure K-bentonite beds to similar beds in coeval stratigraphic sections in eastern North America is tentative.

INTRODUCTION

Altered volcanic ash beds or K-bentonites occur throughout much of the Middle Ordovician stratigraphic sequence of eastern North America (Fox and Grant, 1944; Huffman, 1945; Brun and Chagnon, 1979; Cisne and others, 1982; Huff, 1983; Cullen-Lollis and Huff, 1986; Kolata and others, 1986; Bergström, 1989). These beds are potentially important stratigraphic marker beds because ash falls occur over extremely short periods of geologic time and may have a wide geographical distribution. For decades, stratigraphers have recognized the enormous stratigraphic potential of K-bentonite beds in regional and cosmopolitan correlation of Lower Paleozoic stratigraphic sequences (Kay, 1935; Templeton and Willman, 1963). However, Ordovician K-bentonite beds have been used as marker beds with only limited success because: 1) comprehensive stratigraphic studies documenting the position of these beds within the complex facies of many stratigraphic sequences are yet to be completed; 2) K-bentonite beds generally display similar petrologic and mineralogic features and may vary in thickness, color, texture, and general appearance laterally within a single bed or between two or more beds (Huff, 1983); 3) many K-bentonite beds lack lateral continuity within and between various regions because of nondeposition, erosion, reworking by physical and biological processes, and chemical alteration by diagenesis and metamorphism (Kolata and others, 1986).

Many studies (e.g. Kolata and others, 1986; Samson and others, 1988; Bergström, 1989) have utilized biostratigraphic correlation, chemical fingerprinting from bulk chemical analysis, or mineral chemistry of phenocrysts as approaches to overcome these difficulties. These approaches have had varying degrees of success in the regional correlation of Ordovician K-bentonite beds. One of the successful methods is chemical fingerprinting from bulk chemical analysis. Kolata and others (1986) have successfully correlated two widespread Middle Ordovician K-bentonite beds approximately 900 kilometers from St. Paul, Minnesota to Cape Girardeau, Missouri. Widespread application of this method has been hindered by, among other factors, a lack of comprehensive stratigraphic studies documenting the occurrence of K-bentonite beds in the subcrop of many Ordovician stratigraphic sequences throughout eastern North America (W. Huff, pers. commun., 1990).

The goal of this study is to provide additional information on the stratigraphic position, depositional environment, and physical, mineralogical, and petrological characteristics of previously unrecognized, impure K-bentonite beds from Middle Ordovician rocks in the subsurface of northern Kentucky and southwestern Ohio. In addition, we have tentatively correlated these beds to coeval Middle Ordovician K-bentonite beds in eastern North America utilizing available chronostratigraphic frameworks.

METHODS

The impure K-bentonite beds discussed in this study were initially discovered in the course of routine wireline geophysical logging of Ohio Geological Survey core 2626 in Highland County, Ohio (Figures 1 and 2). A single bed represented by a distinctive wireline log signature was observed (Figure 2). Subsequent lithologic description and X-ray diffraction analysis revealed a concentration of euhedral and subhedral biotite grains and the presence of mixed-layer illite-smectite (Figure 2). Additional investigations, including lithologic description and wireline geophysical logging, led to the recognition of impure K-bentonite beds in nine continuous cores from northern Kentucky and southwestern Ohio (Figure 1).

Preliminary identification of potential K-bentonite beds was accomplished using color, induration, wireline-geophysical-log signatures (when available), and the presence of biotite. The impure K-bentonites were classified according to the shale classification of Potter and others (1980). Rock colors were determined using the Rock-Color Chart (Goddard and others, 1951).

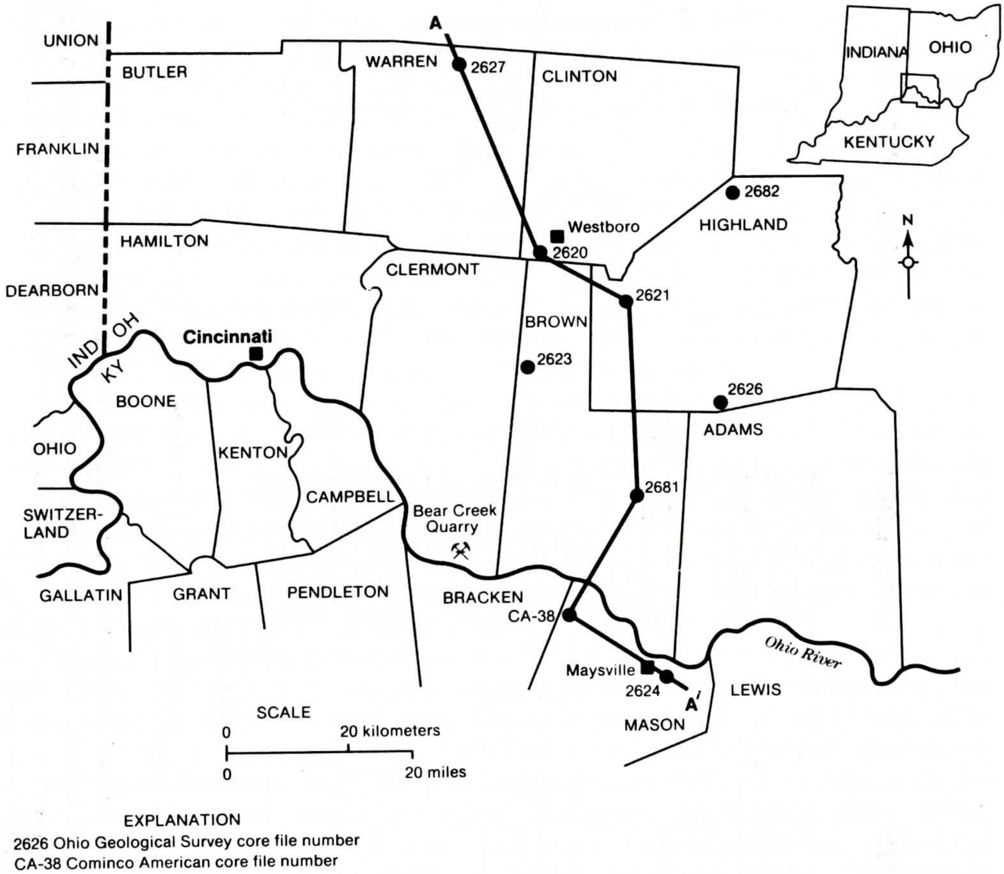

Figure 1. Study area, core locations, and location of cross section A-A' of Figure 3.

Suspected impure K-bentonite beds were positively identified using X-ray diffraction, light-and heavy-mineral analyses, and microscopic examination of 41 samples taken from the impure K-bentonite beds and the overlying and underlying terrigenous shale beds (Table 1). A typical sample consisted of a 1/4 to 1/2 vertical split of BCQ (3.65 cm or 1 3/8 inches) or NCQ (4.76 cm or 1 7/8 inches) diameter core. Subsamples weighing 2 to 3 grams were crushed to less-than-6 mm (1/4 inches) in diameter, then pulverized to less-than 60-mesh in a ceramic mill. A small split of the minus-60-mesh sample was pulverized with a mortar and pestle to minus-200-mesh. Random powder packs of the minus-200-mesh-sample were used for whole-rock X-ray diffraction analysis using Cu K alpha radiation. Carbonates were removed from the 60-mesh samples by adding 1 molar acetic acid buffered with sodium acetate to pH 4.4. The samples were washed in distilled water, using a centrifuge, two or three times or until noticeable flocculation ceased. Each of the washed samples was wet sieved using a 200-mesh screen. The coarse fraction was further disaggregated in an ultrasonic bath for 15 minutes and wet

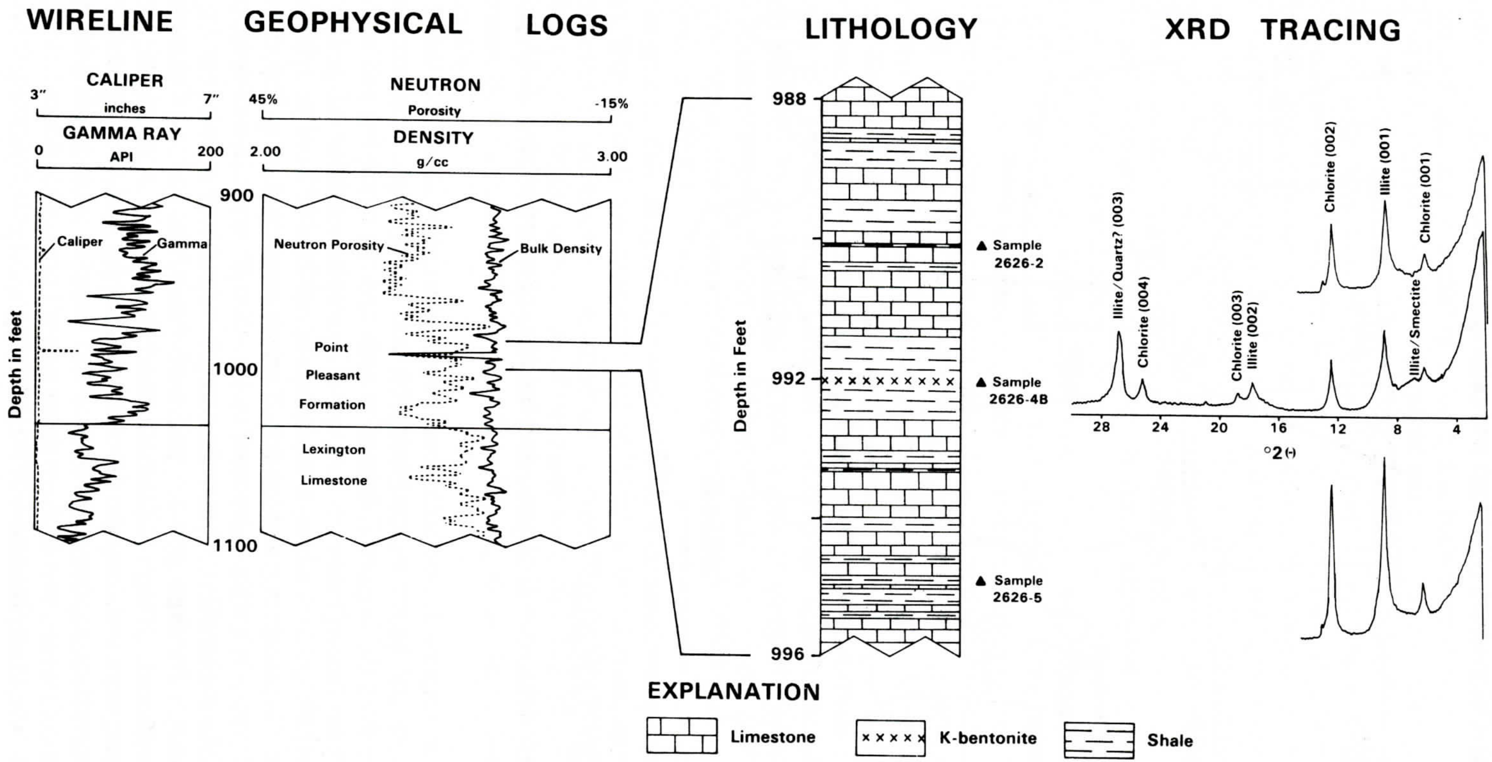

Figure 2. Geophysical logs, lithology, and X-ray diffraction patterns illustrating the impure K-bentonite bed(s) present in Ohio Geological Survey core 2626 in Highland County, Ohio. Note the distinctive kick on the caliper, gamma ray, neutron porosity, and bulk density logs at the position of the impure K-bentonite at 992 feet (301.5 meters), and the presence of interstratified mixed-layer illite-smectite in sample 2626-4B

sieved again. The fine material (less-than-200-mesh) was dispersed with sodium hexametaphosphate in 400 ml of distilled water.

The less-than-2-micrometer clay was extracted from the suspended sample using standard sedimentation techniques described by Folk (1965). Excess water from 20 ml of the less-than-2-micrometer suspension was removed using a millipore filter apparatus and vacuum. Clay samples were washed with 1 N magnesium chloride solution and distilled water using the millipore filter and vacuum. A small amount of clay was removed from the filter and spread onto a glass slide to produce a smear-mount sample for powder X-ray diffraction analysis (Theisen and Harward, 1962). Two smear mounts were made for each sample. The remaining clay was removed from the millipore filter and boiled for 1/2 hour in 2N hydrochloric acid in order to remove chlorite. Smear mounts also were made of this material. In addition, the less-than-0.5-micrometer clay fraction was obtained from two magnesium-saturated samples, one a suspected impure K-bentonite and the other a marine terrigenous shale. A centrifuge was used to accelerate the sedimentation rates of the extremely fine clays.

Thin sections were made for 14 selected samples (Table 1) using a dry grinding technique described by Martin and others (1979). For examination of the coarser minerals, small fragments of several representative samples were soaked in distilled water overnight then alternately crushed with a rubber pestle and subjected to an ultrasonic bath until disaggregated. Heavy minerals in the size range of 80 by 325 mesh were separated from the light minerals using bromoform with a specific gravity of 2.86. Temporary grain mounts were made of both the heavy and light fraction for selected samples (Table 1).

The lensing and truncation of the interbedded limestone and shale beds which characterize the Lexington Limestone and the Point Pleasant Formation make the lateral correlation of individual beds difficult. Sweet and others (1974), Stith (1986), and Schumacher and others (1991) have resolved this problem through the use of shale-percentage logs, which characterize the vertical variations in the percentage of shale of Middle and Upper Ordovician stratigraphic units. These logs are an unsmoothed record of shale thickness stated as a percentage of the total thickness of successive 3-foot intervals (Sweet and others, 1974). The contact between the limestone-dominant Lexington Limestone and the shale-dominant Point Pleasant Formation was determined utilizing shale-percentage logs. This contact served as the datum for determining the stratigraphic position of the various impure K-bentonite beds of this study (Figure 3).

MINERALOGICAL AND PETROLOGICAL RESULTS

Petrographic analysis of thin sections and grain mounts of light and heavy minerals and X-ray diffraction analysis of the samples revealed distinct mineralogical differences between the K-bentonites and marine shales. Light minerals found in the K-bentonite beds include microcline, plagioclase, untwinned feldspar, quartz, mica, and kaolinite. Kaolinite occurs as flakes of low birefringence less-than-3-micrometers long on the surface of the feldspar. In some cases, the kaolinite alteration appears to follow cleavage traces. Kaolinite was positively identified in most of the K-bentonite beds by X-ray diffraction (Table 2). Feldspars generally have moderate amounts of inclusions 1-to 3-micrometers in diameter that give the feldspars a slightly "dirty" appearance. Feldspar edges are corroded but appear to have little if any alteration products. A single zoned

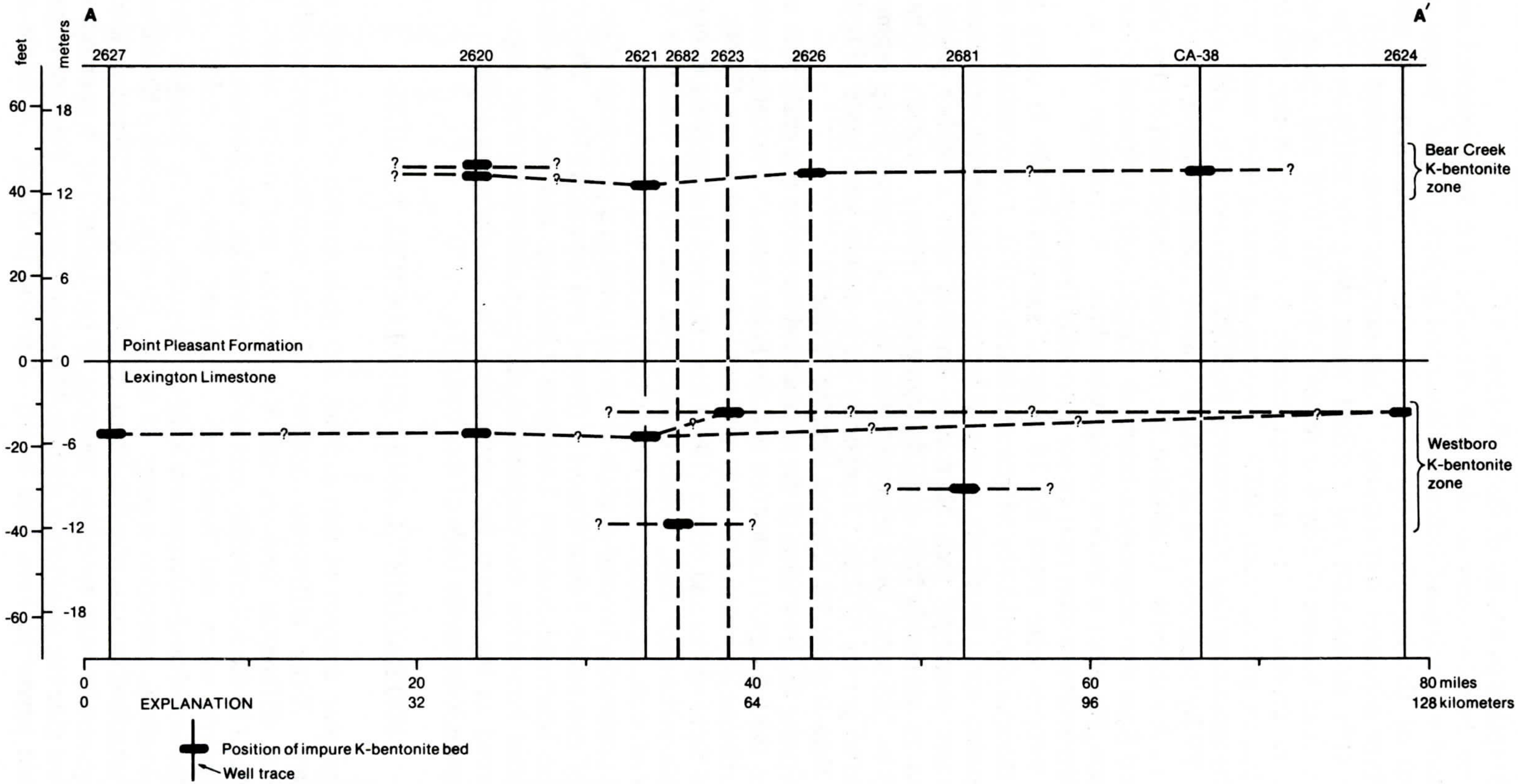

Figure 3. Cross section A-A' showing stratigraphic position of impure K-bentonite beds of the Westboro and the Bear Creek K-bentonite zones relative to the Lexington-Point Pleasant Formational contact. See Figure 1 for location of cross section. Dashed well traces depict the extrapolated position of those cores drilled adjacent to cross section A-A'.

Table 1. Summary of data used and analyses performed on the impure K-bentonite beds and overlying and underlying shale beds of this study.

Core[1]	Location	Core diameter cm (inches)	Stratigraphic unit	Sample number (B = bentonite bed)	Sample interval in feet[2]	Analyses performed[3]
OGS 2620	Clinton Co., Ohio	4.76 (1.875)	Pt. Pleasant Formation Lexington Limestone	2620-1B	744.40 - 744.49	HSD, XRD
				2620-2B	756.52 - 756.57	HSD
				2620-3B	756.57 - 756.67	HSD
				2620-4B	756.96 - 757.00	HSD
				2620-5B	757.01 - 757.03	HSD
				2620-6B	757.08 - 757.10	HSD
				2620-7B	792.65 - 792.70	HSD
				2620-8B	817.91 - 817.94	HSD
				2620-9B	818.05 - 818.10	HSD
OGS 2621	Highland Co., Ohio	4.76 (1.875)	Pt. Pleasant Formation Lexington Limestone	2621-1	756.24 - 756.30	HSD, XRD, TS
				2621-2B	757.15 - 757.25	HSD, XRD, TS
				2621-3	757.95 - 758.00	HSD, XRD
				2621-4	817.36 - 817.40	HSD, XRD, TS
				2621-5B	817.50 - 817.53	HSD, XRD, TS
				2621-6B	817.61 - 817.65	HSD, XRD
				2621-7	817.93 - 817.97	HSD, XRD, TS
OGS 2623	Brown Co., Ohio	4.76 (1.875)	Lexington Limestone	2623-1B	591.20 - 591.25	HSD, XRD
				2623-2	593.90 - 593.95	HSD, XRD
OGS 2624	Mason Co., Kentucky	4.76 (1.875)	Lexington Limestone	2624-1	791.00 - 791.03	HSD, XRD
				2624-2B	792.35 - 792.40	HSD, XRD, LH
				2624-3	792.85 - 792.90	HSD, XRD
OGS 2626	Highland Co., Ohio	4.76 (1.875)	Pt. Pleasant Formation	2626-1	989.90 - 990.00[4]	HSD, XRD, GPL
				2626-2	989.00 - 989.10	HSD, XRD, GPL
				2626-3	992.00 - 992.10	HSD, XRD, TS, GPL
				2626-4B	992.20 - 992.30	HSD, XRD, TS, GPL
				2626-5	994.95 - 995.05	HSD, XRD, GPL
				2626-6	996.70 - 996.80	HSD, XRD, TS, GPL
OGS 2627	Warren Co., Ohio	4.76 (1.875)	Pt. Pleasant Formation Lexington Limestone	2627-1	860.30 - 860.36	HSD, XRD, TS, GPL
				2627-2B?	860.75 - 860.80	HSD, XRD, TS, GPL
				2627-3	861.42 - 861.45	HSD, XRD, GPL, LH
				2627-4	944.50 - 944.52	HSD, XRD, GPL
				2627-5B	945.05 - 945.30	HSD, XRD, TS, GPL
				2627-6	945.50 - 945.60	HSD, XRD, TS, GPL, LH
OGS 2681	Brown Co., Ohio	3.65 (1.375)	Lexington Limestone	2681-1	738.80 - 738.90	HSD, XRD
				2681-2B	739.05 - 739.10	HSD, XRD
				2681-3	739.95 - 740.00	HSD, XRD
OGS 2682	Highland Co., Ohio	3.65 (1.375)	Lexington Limestone	2682-1	1136.50 - 1136.52	HSD, XRD
				2682-2B	1136.85 - 1136.90	HSD, XRD
				2682-3B	1136.90 - 1136.95	HSD, XRD, TS, LH
				2682-4B	1136.95 - 1137.05	HSD, XRD, TS
				2682-5	1138.26 - 1138.32	HSD, XRD
CA-38	Mason Co., Kentucky	3.65 (1.375)	Pt. Pleasant Formation	Not sampled		HSD

[1]OGS - Ohio Geological Survey core file number; CA-38 - core drilled by Cominco American Inc., stored at The Ohio State University.
[2]Metric equivalents not given because the Ohio Geological Survey uses English units for core measurements.
[3]HSD - hand sample description
XRD - x-ray diffraction
TS - thin section
LH - light and heavy minerals
GPL - geophysical log suite
[4]Error in sample numbering.

volcanic feldspar was identified in one sample.

Heavy minerals found in the impure K-bentonites include euhedral to anhedral biotite, idiomorphic zircon and apatite, and pyrite. Biotite may be associated with hematite, and rarely magnetite. Pyrite occurs as framboids less-than-10-micrometers in diameter, as irregular replacements, and as tiny grains and inclusions less-than-2-micrometers in diameter. Pyrite is more abundant than magnetite in all thin sections examined.

Samples recovered from several centimeters above and below the suspected K-bentonites are considered marine shales of nonvolcanic origin. They are more indurated and contain more silt-size quartz and calcite than the impure K-bentonites (Table 2).

Petrographic examination reveals the presence of dolomite rhombs up to 180 micrometers in diameter in most samples. Degraded as well as fresh biotite can be found in small quantities in many of the marine shales. The most abundant light minerals found in the these shales are quartz and feldspars. Microcline and

Table 2. X-ray mineralogy of shale and limestone samples. Peak intensities are shown only for comparisons between samples listed in this table. ND = not determined; M = major mineral component; T = minor or trace mineral component; B = impure bentonite; I/S = illite/smectite mixed-layer clay.

Sample number	Nonclay mineralogy				Clay mineralogy			
	Quartz	Calcite	Dolomite	Pyrite	I/S	Illite	Chlorite	Kaolinite
2621-1	100	30	13	8		M	T	
2621-2B	100	55	11	8	T	M	T	
2621-3	100	34	16	10		M	T	
2621-4	86	84	9	7	?	M	T	
2621-5B	27	100	0	5	M	T	T	T
2621-6B	48	40	10	8	M	T	T	T
2621-7	81	100	29	5		M	T	
2623-1B	100	30	13	8	T	M	T	T
2623-2	100	10	15	12	?	M	T	?
2624-1	100	34	34	10		M	T	T
2624-2B	47	17	8	8	M	T	T	
2624-3	72	100	30	7		M	T	
2626-1	100	39	25	9		M	T	ND
2626-2	100	32	30	8		M	T	ND
2626-3	100	16	14	7		M	T	ND
2626-4B	ND	ND	ND	ND	ND	ND	ND	ND
2626-5	100	17	0	8	T	M	T	
2626-6	100	11	12	11		M	T	ND
2626-7	100	15	11	9		M	T	ND
2627-1	100	36	10	9		M	T	?
2627-2B	100	68	13	7	?	M	T	
2627-3	100	67	19	7		M	T	
2627-4	90	100	19	7		M	T	
2627-5B	41	90	8	6	M	T	T	T
2627-6	100	100	23	5		M	T	
2681	41	100	6	6	?	M	T	
2681-2B	38	10	8	7	M	T	T	T
2681-3	100	37	22	7		M	T	
2682-1	91	46	9	7		M	T	?
2682-2B	82	52	9	6	M	T	T	T
2682-3B	ND	ND	ND	ND	ND	ND	ND	ND
2682-4B	36	43	0	10	M	T	T	T
2682-5	100	69	30	7		M	T	

plagioclase were identified but make up only small portions of the rock.

Few heavy minerals are present in the marine shales. The most abundant component of sample 2627-6 is dolomite rhombs which were not dissolved by the warmed, buffered acetic acid. Pyrite, rounded collophane, and tourmaline grains and some quartz with abundant magnetite inclusions make up the majority of the heavy-mineral suite.

When bentonites are buried, diagenetic and low-grade metamorphic recrystallization may first transform the smectite into mudstones and shales rich in mixed-layer illite-smectite (Roberts and Merriman, 1990). Eventually, under the right conditions the illite-smectite may be transformed into a pure illite (Huff and Türkmenoglu, 1981). Thus, at some stage the original volcanic ash may contain a mixed-layer clay and a characteristic non-clay, volcanic mineral assemblage. If the suspected K-bentonites are truly of volcanic origin with little or no contamination from nonvolcanic sources, the expected illite produced from low-temperature diagenesis of the volcanic ash is the 1Md polytype.

X-ray diffraction analysis was performed on the less-than-2-micrometer size fraction for 32 samples and on the less-than-0.5-micrometer size for selected

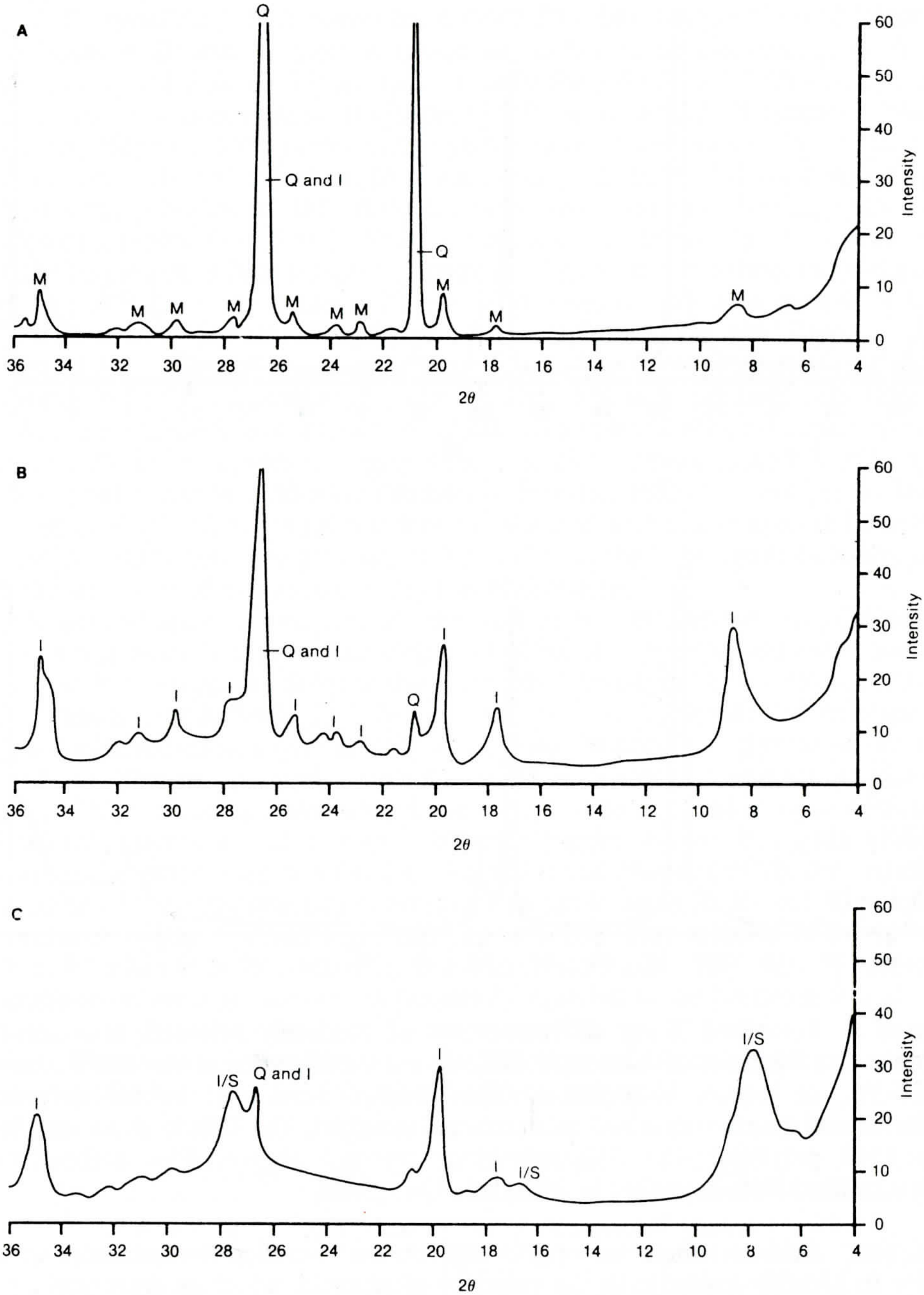

Figure 4. Smoothed X-ray diffractograms of randomly oriented, less-than-2-micron size fraction of A, muscovite from a pegmatite; B, normal marine shale, sample 2624-3, and C, impure K-bentonite, sample 2624-2B. All samples were acid treated. Quartz was added to the muscovite (A) as an internal standard. Many strong unique reflections for 2M muscovite occur between 20 and 35 degrees two theta. A and B mostly contain this polytype. However, C has strong, unique muscovite peaks only at 19.8 degrees (4.48 Å) and 35 degrees (2.56 Å) indicating 1Md illite polytype. I/S= ordered (IM) interstratified illite-smectite, I= illite, M=muscovite, and Q=quartz.

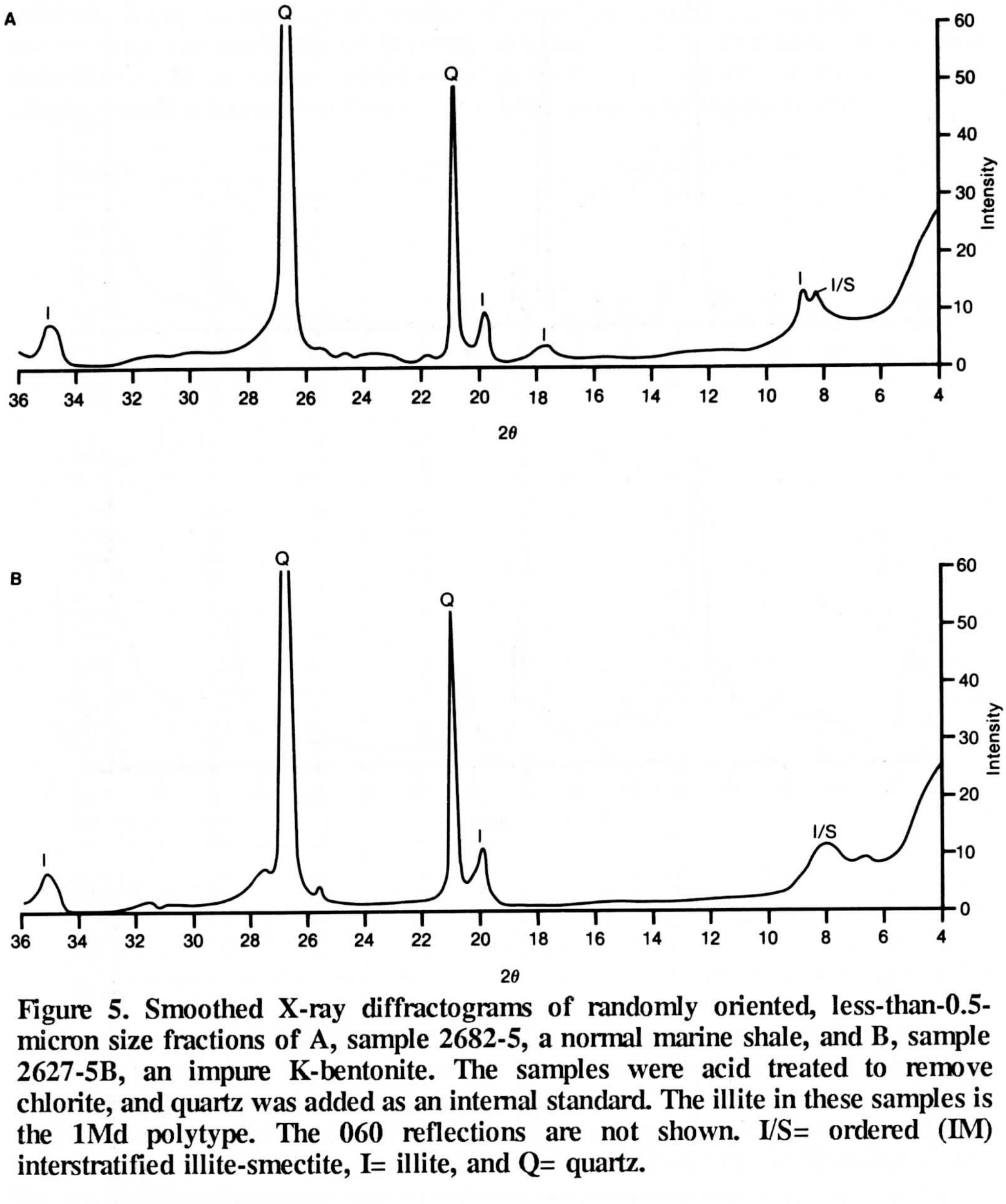

Figure 5. Smoothed X-ray diffractograms of randomly oriented, less-than-0.5-micron size fractions of A, sample 2682-5, a normal marine shale, and B, sample 2627-5B, an impure K-bentonite. The samples were acid treated to remove chlorite, and quartz was added as an internal standard. The illite in these samples is the 1Md polytype. The 060 reflections are not shown. I/S= ordered (IM) interstratified illite-smectite, I= illite, and Q= quartz.

samples. Acid treatment was performed on the less-than-2-micrometer size in order to identify kaolinite in the presence of chlorite, which is destroyed by the acid.

The less-than-0.5-micrometer size fraction of sample 2682-5, a marine shale, and sample 2627-5B, a K-bentonite, were X-rayed in a random powder mount with a small amount of quartz added to the sample to serve as an internal standard. The 10 Å mineral in sample 2682-5 and sample 2627-5B has 060 reflections of 1.501 and 1.499 , for Cu K alpha radiation, respectively. This peak is routinely used to determine the nature of the octahedral sheet of illite, and in both samples (2682-5 and 2627-5B) the 060 reflection indicates dioctahedral illite (Starkey and others, 1985).

Well-crystallized, high-temperature (2M) illite has distinct X-ray diffraction peaks between 20 and 35 degrees two theta, including well developed peaks at about 4.47 Å and 2.56 Å (Figure 4A). If only the 4.47 Å and 2.56 Å peaks are present, the low-temperature polytype (1Md) is indicated (Reynolds, 1980). Velde and Hower (1963) report that high-temperature 2M muscovite, if present, occurs in the coarse size fraction of marine shales. Because 2M muscovite peaks can mask the presence of 1Md illite, the very fine less-than-0.5-micrometer size fraction of samples 2682-5 and 2627-5B were X-rayed (Figure 5). 1Md illite was found to be present in both samples. A peak at 17.8 degrees two theta and a slightly more "jagged" base line between 20 and 35 degrees two theta in sample 2682-5 suggests that some 2M illite may be present in addition to the 1Md illite.

In addition, the diffraction pattern of the less-than-2-micrometer size fraction of sample 2624-2B (K-bentonite) suggests that 1Md illite predominates (Figure 4C). In sample 2624-3, a marine shale, additional non-basal illite reflections appear indicating 2M mica is present (Figure 4B). The distinct peak at 2.99 Å in sample 2624-3 is not found in 1Md illite (Yoder and Eugster, 1955). Hence, in the less-than-2-micrometer clay, 2M mica is found in the marine shale, as expected, but not in the K-bentonite bed. The presence of 1 Md illite in the K-bentonite bed indicates the illite was formed in a low-temperature environment.

X-ray diffraction analysis of the less-than-2-micrometer clay minerals indicates that most of the samples suspected of being impure K-bentonites contain an ordered illite-smectite similar to those described by Reynolds and Hower (1970) and Reynolds (1980, Table 3). The primary peak for ordered illite-smectite in these samples occurs at approximately 13 Å when the sample is glycolated (Figure 6). All samples in this study contain ordered illite-smectite and are similar to profiles of illite-smectite (IM) ordered interstratification depicted in figures 9E, 9F, and 9G of Reynolds and Hower (1970) or Figure 4.8 of Reynolds (1980). Following the procedures described by Reynolds and Hower (1970), the smectite component of the illite-smectite is estimated to range between 20 and 30 percent. The volcanic mineral assemblage, and illite-smectite clay mineral found in this study are distinct from the overlying and underlying rocks. This distinct volcanic mineral/clay mineral assemblage is frequently reported in the literature and is

Table 3. Summary of the thickness and lithology of impure K-bentonite beds within the Westboro and Bear Creek K-bentonite zones.

OGS[1] core number: sample number	Thickness in meters (feet)	Lithology	K-bentonite zone
2620: 2B-6B	0.11 (0.35)	Fossiliferous claystone	Westboro
2620: 8B, 9B	0.46 (1.50)	Biotite-rich rhythmite Burrow-fill	Bear Creek
2621: 2B	0.15 (0.50)	Biotite-rich rhythmite Burrow-fill	Bear Creek
2621: 5B, 6B	0.09 (0.30)	Fossiliferous claystone	Westboro
2623: 1B	0.03 (0.10)	Fossiliferous claystone	Westboro
2624: 2B	0.15 (0.50)	Fossiliferous claystone	Westoro
2626: 4B	0.21 (0.70)	Biotite-rich rhythmite	Bear Creek
2627: 5B	0.06 (0.20)	Fossiliferous claystone	Westboro
2681: 2B	0.09 (0.30)	Fossiliferous claystone	Westboro
2682: 2B, 4B	0.05 (0.15)	Fossiliferous claystone	Westboro
CA-38[2] not sampled	0.01 (0.05)	Burrow-fill	Bear Creek

[1]Ohio Geological Survey.
[2]Cominco American, Inc.

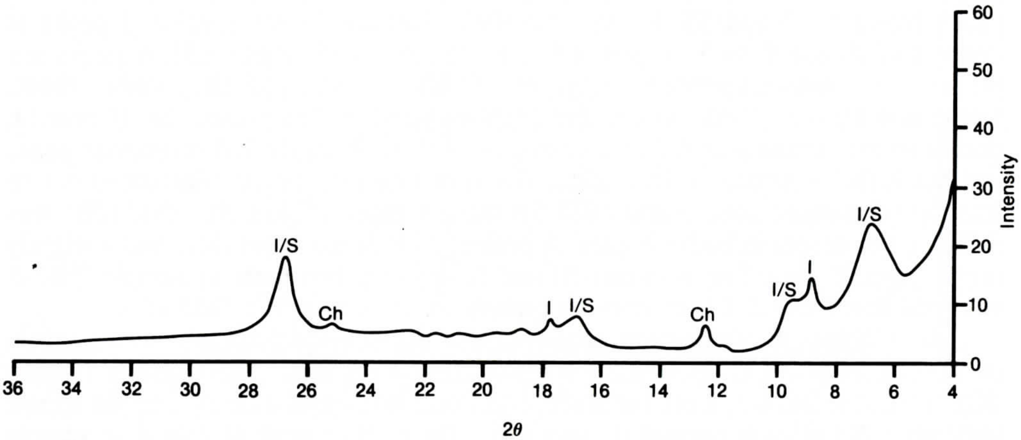

Figure 6. Smoothed X-ray diffractogram of oriented, less-than-2-micron size fraction of sample 2624-2B, an impure K-bentonite. The illite-smectite mixed-layer clay contains approximately 25% expandable clay. The sample was magnesium saturated and ethylene glycolated prior to analysis. I/S = ordered (IM) interstratified illite-smectite (Reynolds, 1980), Ch = chlorite, and I = illite.

universally accepted as strong proof of a volcanic origin. The illite-smectite bearing rocks in this study are considered to be volcanic in origin, but because chlorite, which is considered terrigenous, occurs in all samples X-rayed, the rocks are considered to be impure K-bentonites.

K-BENTONITE STRATIGRAPHY

Previous Work

Numerous K-bentonite beds have been recognized within the Middle and Upper Ordovician stratigraphic sequence of northern Kentucky and southwestern Ohio. A minimum of five to a maximum of 11 individual K-bentonite beds have been reported from the upper 35.0 meters (115 feet) of the Black River Group and the basal 15.2 meters (50 feet) of the Lexington Limestone in 16 cores drilled in northern Kentucky and southwestern Ohio (Sweet and others, 1974; Stith, 1979 and 1986; Huff, 1983; Shrake and others, 1990)(Figure 7). Two impure K-bentonite beds were reported from the Lexington Limestone and Point Pleasant Formation in eight cores drilled in northern Kentucky and southwestern Ohio (Schumacher and Carlton, 1989). W. C. Sweet (pers. commun.) and Sweet (1987) indicate the presence of a possible K-bentonite bed approximately 22.8 meters (75 feet) below the top of the Point Pleasant Formation in the Bear Creek quarry in Clermont County, Ohio (Figures 1 and 7). Shrake and others (1990), utilizing the preliminary results of Schumacher and Carlton (1989), recognized a single K-bentonite bed from the upper Lexington Limestone in Ohio Geological Survey core 2627 in Warren County, Ohio. Caster and Kjellesvig-Waering (1964) reported a K-bentonite bed 1.7 meters (5.6 feet) below the top of the informal Cincinnati group in Adams County, Ohio (Figure 7). Currently, this bed is not considered a K-bentonite (W. Huff, pers. commun., 1990). However, the stratigraphic interval containing this bed has not been restudied to confirm or deny the existence of

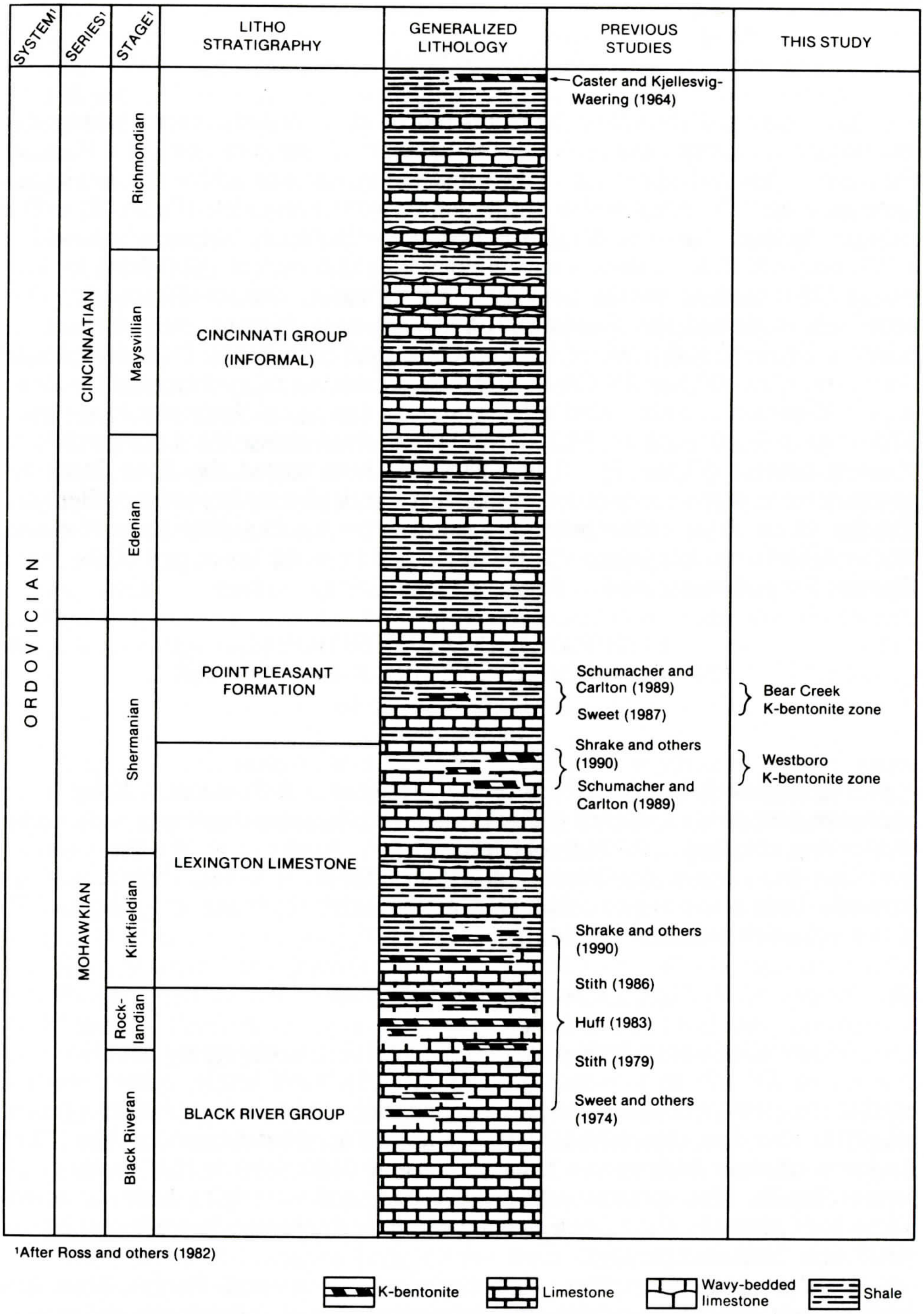

Figure 7. Schematic diagram summarizing previous and current investigations, chronostratigraphy, and lithostratigraphy of Middle and Upper Ordovician K-bentonites of northern Kentucky and southwestern Ohio.

this K-bentonite bed.

Present Work

The impure K-bentonite beds discussed in this study occur within the interbedded limestones and shales of the Lexington Limestone and Point Pleasant Formation. These beds occur as two discrete zones: one within the Lexington Limestone and the other within the Point Pleasant Formation (Figure 3). The stratigraphically lower zone consists of two to four impure K-bentonite beds within a 7.9-meter-(26.0 foot)-thick zone ranging from 3.6 meters (12.0 feet) to 11.6 meters (38.0 feet) below the Lexington-Point Pleasant contact (Figure 3). This zone is here named the Westboro K-bentonite zone because these impure K-bentonites were first discovered in Ohio Geological Survey core 2620 drilled near Westboro, Ohio (Figure 1). The second zone consists of two or possibly three impure K-bentonite beds within a 1.2 meter-(4.0 foot)-thick interval ranging from 12.5 meters (41.0 feet) to 13.7 meters (45.0 feet) above the Lexington-Point Pleasant contact (Figure 3). This interval is here named the Bear Creek K-bentonite zone was once exposed in the Bear Creek Quarry in southern Clermont County, Ohio. The authors wish to acknowledge that Drs. Stig Bergström and Walter Sweet first recognized a K-bentonite bed from the lower part of the Point Pleasant Formation exposed in the Bear Creek Quarry.

LITHOLOGY AND DEPOSITIONAL ENVIRONMENT OF IMPURE K-BENTONITES

Introduction

The impure K-bentonite beds of the Westboro and Bear Creek K-bentonite zones are generally olive-gray (5Y 4/1), hard, argillaceous claystones with sparse to abundant euhedral to subhedral biotite grains. In hand sample, the Westboro K-bentonite beds are a fossiliferous claystone (Table 3). The Bear Creek K-bentonite beds occur as two lithologies: a biotite-rich rhythmite and a burrow fill with biotite-rich sediment (Table 3).

Fossiliferous Claystone

Impure K-bentonite beds classified as fossiliferous claystones are olive gray (5Y 4/1 or 5Y 3/2) to brownish gray (5 YR 4/1), hard, brittle, calcareous, and argillaceous. They contain sparse to abundant subhedral to euhedral biotite grains, sparse to abundant thin-valved brachiopods, and rare calcareous laminae. The thickness of these beds ranges from 0.03 meter (0.10 foot) to 0.15 meter (0.50 foot)(Table 3). The contacts with the overlying and underlying beds are sharp. These beds generally slake when wetted and rarely display crude grading of biotite (OGS core 2682, sample 2B).

Hand-sample description, thin-section analysis, and X-ray diffraction analysis indicate that these beds are a mixture of altered volcanic ash and marine sediments. These beds are interpreted to have originated from mixing of volcanic ash with marine sediments through the action of storm sedimentation. We envision the following sequence of events. Initially, volcanic ash was deposited over widespread areas in the epeiric sea that covered the area of northern

Kentucky and southwestern Ohio in the Middle Ordovician. Marine sedimentation along with colonization by opportunistic brachiopod species (Martin, 1975) produced an overlying deposit of shale with abundant brachiopods. This deposit, along with the underlying volcanic ash, was eroded, resuspended, transported, and redeposited by storm-generated unidirectional and oscillatory currents. The resulting beds of impure K-bentonite will display a sharp lower contact and consist of a crudely graded mixture of whole opportunistic-brachiopod valves, altered volcanic ash, and marine shale and limestone. The impure K-bentonite beds grade into overlying marine shale and limestone beds as observed in some fossiliferous claystones of this study.

These deposits are similar to the storm-generated sequences described by Kreisa (1981) and Meyer and others (1981). These sequences are characterized by a sharp, erosional lower contact overlain by a graded sequence of coarse- to fine-grained skeletal fragments and argillaceous material which in turn grade into overlying fine-grained sediments.

Lithofacies of the Lexington Limestone exposed in central Kentucky exhibit considerable evidence of storm sedimentation. Overturned stromatoporoids, calcarenite beds, rippled beds, broken bryozoan colonies, crude graded bedding, rip-up clasts, and scour surfaces have been reported within many of the subtidal lithofacies within the Lexington Limestone (Cressman, 1973; Ettensohn and others, 1986). Similar sedimentary structures have been observed in the Lexington Limestone and the Point Pleasant Formation in the course of core description in this study.

Biotite-rich Rhythmites

Impure K-bentonite beds classified as biotite-rich rhythmites contain sparse to abundant subhedral to euhedral biotite grains which are concentrated within alternating carbonate and clastic beds or laminae. Carbonate beds or laminae are medium gray (N5) to olive gray (5Y 4/1), generally fine grained, and crudely graded. They consist of skeletal fragments, calcareous cements, argillaceous material, and biotite. The thickness of these beds range from 0.15 meter (0.50 foot) to 0.46 meter (1.50 feet) (Table 3). The contact with the underlying clastic unit is sharp, and the upper contact is gradational to sharp. Clastic beds or laminae are olive gray (5Y 4/1), hard, well indurated, calcareous, argillaceous, and sparsely fossiliferous. These beds contain sparse to abundant biotite grains. The contact with the underlying carbonate unit is gradational to sharp, and the upper contact is sharp.

This lithology is well illustrated by the rocks between 229.50 meters (755.06 feet) and 230.50 meters (758.35 feet) in Ohio Geological Survey core 2620 (Figure 8). The rhythmites within this interval represent one or more volcanic ash beds which were subjected to repeated exhumation, mixing with normal marine sediments, and redeposition by storm-generated processes.

The impure K-bentonite beds of the Bear Creek K-bentonite zone were deposited at approximately the same paleogeographic position as those of the Westboro K-bentonite zone, within the paleostorm track for the southern hemisphere throughout the Middle Ordovician (Kreisa, 1981). As with the Westboro K-bentonite beds, the sedimentary sequences and structures exhibited by biotite-rich rhythmites are similar to those seen in storm-dominated sedimentary sequences. Sedimentary features consisting of sharp erosional(?) lower contacts and crude graded bedding support this interpretation.

Burrow-fill Lithology

This lithology generally occurs in conjunction with biotite-rich rhythmites in the Bear Creek K-bentonite zone. The burrow-filling sediments are olive gray (5Y 4/1), hard, argillaceous, and calcareous. Sparse to abundant biotite grains are concentrated within burrows. Burrows range in thickness from 0.01 meter (0.05 foot) to 0.03 meter (0.10 foot).

This lithology is well illustrated by the rocks between 230.00 meters (756.70 feet) and 230.10 meters (756.86 feet) in Ohio Geological Survey core 2620 (Figure 8). This lithology is the result of burrowing animals ingesting volcanic ash mixed with normal marine sediments in order to extract nutrients. The organism concentrated the larger size fraction of the sediment through the digestive process and deposited this fraction and fecal material within the burrow.

DISCUSSION

Determination of the origin of the micaceous clay minerals in the samples taken adjacent to the K-bentonite beds can shed light on the origin of the shale samples collected in this investigation. The term illite is used as a group name for clay size micaceous material in argillaceous sediments that has a 10 Å basal reflection when analyzed by powder X-ray diffraction. In marine shales, illite may consist of fine- grained muscovite derived from mechanical weathering of igneous and metamorphic rocks or may be the result of low-temperature alteration or neoformation in the marine environment. Paleozoic marine shales generally contain both types (Velde and Hower, 1963). Illite derived from mechanical weathering of high-temperature rocks has a 2M polytype structure, while low-temperature diagenetic illite has 1Md polytype structure (Yoder and Eugster, 1955).

Prior to 1970, many midcontinent Paleozoic rocks identified as K-bentonites were reported to contain randomly interstratified dioctahedral illite and montmorillonite layers (Weaver, 1953; Huff, 1963; Lounsbury and Melhorn, 1964; Mossler and Hayes, 1966). Expandable layers generally constitute 20 to 25 percent of the total structure (Huff, 1963). In 1970, Reynolds and Hower showed what was previously described as random illite-smectite was, in fact, ordered illite-smectite with either an IM or IMII superlattice. The impure K-bentonites found in this study contain ordered illite-smectite with an IM superlattice. The expandable layers range from 20 to 30 percent.

The dominant illite polytype recognized in both the K-bentonites and the enclosing strata in this study appears to be the 1Md polytype, indicating a low temperature of formation, and suggests that the majority of the 10 Å material in the less-than-0.5-micrometer size range in both types of samples is 1Md illite rather than detrital high-temperature muscovite. The 2-micrometer size fraction of the marine rocks appears to have some 2M illite present, whereas X-ray analysis of the K-bentonites show no trace of this high-temperature polytype.

Velde and Hower (1963) also found the 2M polytype more concentrated in the coarser size fractions of marine Paleozoic shales. The non-clay mineral suite, as well as the lithologic distinctiveness from enclosing strata, and blanket like nature of the deposits has been used to prove the volcanic origin of the K-

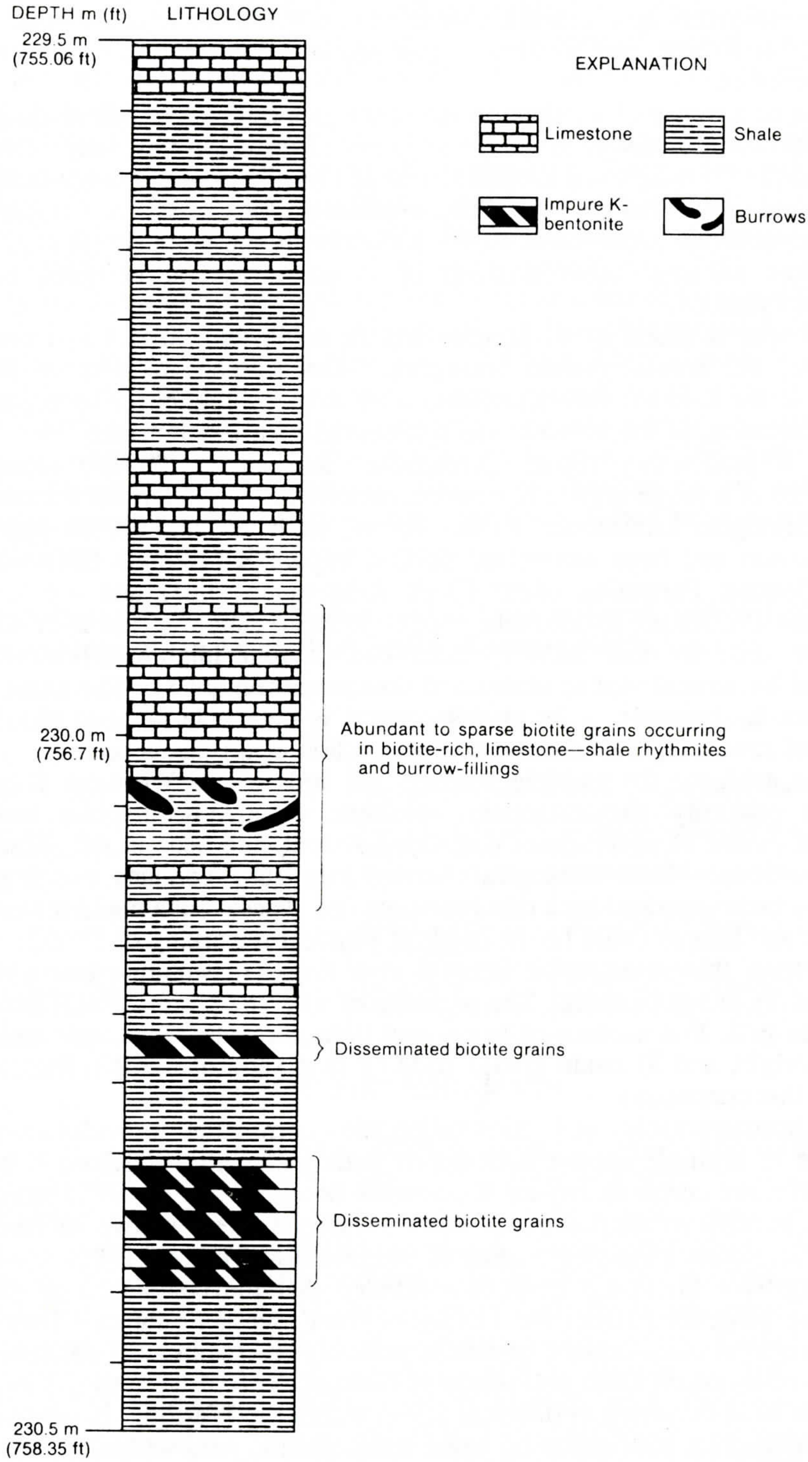

Figure 8. Impure K-bentonite beds in the Bear Creek K-bentonite zone of Ohio Geological Survey core 2620 in Clinton County, Ohio. Note the presence of biotite-rich rhythmites and burrow-fillings.

bentonite-bearing rocks reported in the literature. The non-clay minerals such as idiomorphic zircon, apatite, and biotite in this study are similar to the mineral suites reported for K-bentonites elsewhere, and strongly suggest a volcanic component for these rocks.

The occurrence of kaolinite in the lithologies sampled in this investigation is limited almost exclusively to the K-bentonites identified in this study. Droste and Vitaliano (1973) recognized kaolinite in 10 of 11 samples of the Tioga Bentonite in Indiana and considered the most likely explanation for the presence of kaolinite to be the result of the alteration of albite. Evidence from this study also supports this conclusion, although other methods of formation cannot be ruled out from available evidence.

Chlorite is found in all samples and is, along with illite, considered to be typical of the normal marine sediments. The presence of discrete illite and chlorite in the K-bentonites is probably a result of mixing, either by organisms or storm processes, of the volcanic ash and normal marine sediments.

In a preliminary report, Schumacher and Carlton (1989) reported the occurrence of a single impure K-bentonite bed in the Point Pleasant Formation and in the Lexington Limestone. In this report, we have reconsidered our original interpretation and have recognized several impure K-bentonite beds within the Point Pleasant Formation (Bear Creek K-bentonite zone) and the Lexington Limestone (Westboro K-bentonite zone). Reexamination of the Bear Creek K-bentonite zone in core 2620 revealed two to four impure K-bentonite beds separated by normal marine shales and limestones (Figure 8). The exact number of impure K- bentonite beds in this interval is debatable because physical and biological reworking has mixed the upper portion of this sequence.

The evidence for multiple volcanic ash beds in the Westboro K-bentonite zone is primarily circumstantial. Multiple impure K-bentonite beds were observed in one of seven cores containing K-bentonite beds in the Westboro K-bentonite zone. Ohio Geological Survey core 2682 contains two impure K-bentonite beds separated by a thin limestone bed. These strata could represent two volcanic ash falls or could be the result of physical reworking of a single ash fall. We interpret this stratigraphic interval as a single K-bentonite bed which was reworked by storm currents. The presence of 1) a sharp erosional(?) base of the limestone bed, 2) a mixture of heavy and light minerals of volcanic and normal marine origin, and 3) crude graded bedding at the top of the thin limestone bed support this conclusion.

Lithostratigraphic and biostratigraphic correlation provides supporting evidence of multiple impure K-bentonite beds within the Westboro K-bentonite zone. In some cores, an impure K-bentonite bed occurs at the same stratigraphic position relative to the Lexington-Point Pleasant contact in two or more cores (Figure 3). Biostratigraphic studies of Bergström and others (1990) and Mitchell and Bergström (in press) have demonstrated that the boundaries of diagnostic graptolite biozones parallel the Lexington-Point Pleasant contact. They suggest the Lexington-Point Pleasant contact is probably an isochronous horizon, at least within southwestern Ohio and northern Kentucky. If it is, then the impure K-bentonite beds occurring at different levels within the Westboro K-bentonite zone would be discrete beds assuming equal sedimentation rates within all stratigraphic successions studied. We feel that the contact is an isochronous horizon but determining the exact number of impure K-bentonite beds is beyond the methods used in this study because the impure K-bentonite beds display similar lithologic, mineralogic, and petrologic characteristics. We suggest that additional studies utilizing biostratigraphic correlation, and/or chemical fingerprinting from bulk chemical analysis, and/or mineral chemistry of phenocrysts be undertaken in order to determine the number and lateral correlation of impure K-bentonites in the Westboro K-bentonite zone.

The stratigraphic interval containing the Westboro and Bear Creek K-bentonite zones is referable to the Middle Ordovician, Shermanian Stage of the Mohawkian Series (Sweet, 1979; Bergström and Mitchell, 1986). The position of these impure K-bentonite beds within the Shermanian Stage is approximately the same stratigraphic level as the Haldane, Nassett, Conover, and Calmar K-bentonite beds in the northern Illinois and southeastern Minnesota; one or two unnamed beds in central Kentucky and eastern Tennessee; a number of unnamed K-bentonite beds in central Pennsylvania and New York; and four unnamed beds in southern Quebec (Table 4). However, exact correlation of these K-bentonite beds with K-bentonites from coeval Shermanian stratigraphic sequences cannot be determined because published chronostratigraphic studies on the cores of this study are not available. Chronostratigraphic studies should be undertaken to refine the stratigraphic position of the Westboro and Bear Creek K-bentonite zones relative to coeval Shermanian stratigraphic sequences.

Table 4. Selected K-bentonite beds reported from rocks of Middle Ordovician, Shermanian Stage of the Mohawkian Series of eastern North America.

Region	K-bentonite beds reported	Lithostratigraphic unit	Reference
Northwestern Ohio	Several	Trenton Limestone	Wickstrom and Gray (1988)
Central Kentucky	1	Lexington Limestone	Black and others (1965)
Southern Kentucky	1	Lexington Limestone	Borella and Osborne (1978)
Eastern Tennessee (Knox Co.)	2	Bays Formation	Milici (1973)
Northern Illinois	4	Dunleith Formation	Willman and Kolata (1978)
Southeastern Minnesota	4	Prosser Formation	Sloan (1987)
Central Pennsylvania	2	Salona and Coburn Formations	Berkheiser and Cullen-Lollis (1986)
New York subsurface	Many	Denley Limestone	Rickard (1973)
Mohawk Valley of New York	Many	Utica Shale	Cisne and others (1982)
Quebec, Canada	4	Neuville Formation	Brun and Chagnon (1979)

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NEW DISCOVERY OF *SALTERELLA* IN THE LOWER CAMBRIAN ROME FORMATION, APPALACHIAN FOLD-THRUST BELT, CENTRAL ALABAMA

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ABSTRACT

The Lower Cambrian Rome Formation exposed in a railroad cut at Helena, Shelby County, Alabama, contains abundant specimens of *Salterella*. The locality is at the leading edge of the Helena thrust sheet in the central part of the Appalachian fold-thrust belt in Alabama. Approximately 105 meters of the upper part of the Rome is exposed in the cut, and the fossils occur in three thin beds of limestone in the middle part of the sequence. Specimens collected from the uppermost limestone bed are identified as *Salterella maccullochi* (Murchison), confirming the geographic range of this species in eastern North America from Labrador to Alabama. Because this species is known to occur in the Shady Dolomite and correlative units in other parts of the Appalachians, this discovery potentially links, by means of a common fossil, facies within the Rome Formation and Shady Dolomite.

INTRODUCTION

Abundant specimens of *Salterella* have been discovered in the Lower Cambrian Rome Formation at Helena, Shelby County, Alabama (Figure 1). *Salterella* is geographically widespread and locally very abundant in Lower Cambrian rocks in the Appalachians (Resser, 1938; Yochelson, 1970). Although this genus has been previously reported in Alabama, the localities for the earlier collections are problematical.

Little information is published on the occurrence of *Salterella* in Alabama. McCalley (1897) describes siliceous blue limestone containing abundant *Salterella* in the "Aldrich Limestone" in his report on the valley regions of Alabama. McCalley believed the "Aldrich" to underlie the Lower Cambrian Rome Formation, which he referred to as the "Montevallo (Variegated) Shale and Sandstone." However, Butts (1926, p. 64) recognized that most of the rocks that McCalley referred to as "Aldrich" were in fact Middle to Upper Cambrian Conasauga Formation in the core of an overturned syncline. Consequently, it appears likely that the fossils McCalley (1897) observed were in limestone interbeds in the Rome Formation on the flanks of the structure. Butts (1926, p. 65) mentioned the presence of *Salterella* in the Shady Dolomite in Alabama, citing

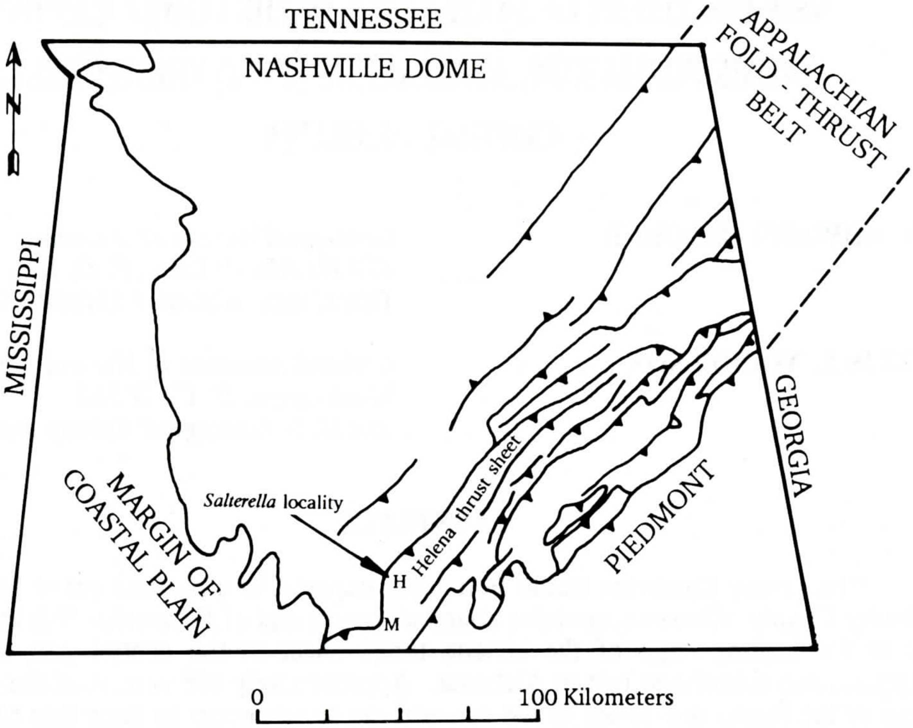

Figure 1. Generalized map of northern Alabama showing major thrust faults within the Appalachian fold-thrust belt and the *Salterella* locality at Helena on the Helena thrust sheet. H - Helena, M - Montevallo.

McCalley (1897, p. 41). Butts also notes that the most extensive outcrops of the Shady known to him were along Beeswax Creek in Shelby County. Independent searches of the exposures in Beeswax Creek by the present authors did not yield any fossils.

The only published documentation of *Salterella* in the state is that of Resser (1938, p. 45, pl. 5, fig. 11). He noted *Salterella* sp. as occurring 1 1/4 miles northwest of Montevallo, Shelby County (Figure 1). His only illustration consists of a view of the weathered surface of a rock, magnified twice natural size. Even though no thin sections have been cut of his material (U. S. National Museum of Natural History specimen USNM 94774), examination of broken surfaces shows it to be conspecific with that illustrated herein.

Resser gave no adequate stratigraphic information and, worse, he neglected additional significant details that were available to him. For one, he did not credit C. D. Walcott and H. McCalley with this discovery, which was made May 4, 1895; presumably McCalley's comments in 1897 about *Salterella* stem from this joint field work. For another, Resser gave imprecise locality data. The material he figured was obtained from a limestone outcrop in a small creek on the south side of a road, 1 1/4 miles west-northwest (not northwest) of Montevallo.

Although the *Salterella* bearing beds discovered by McCalley and Walcott have not been located, examination of outcrops west- northwest of Montevallo by the senior author indicates a stratigraphic sequence similar to that exposed at Helena, suggesting that the fossils occur in essentially the same stratigraphic position. In the early 1970's Charles Cressler, U.S. Geological Survey Water

Resources Division (now retired), collected *Salterella* from outcrops in the Montevallo area. These outcrops could have been in the same general area of the McCalley and Walcott locality, but the exact location of the outcrops he visited has been lost.

In 1990 while doing detailed mapping at Helena in the Cahaba Valley of Shelby County, Alabama (Osborne, in press), Osborne discovered abundant small cone-shaped fossils in limestone beds within the Rome Formation. A sample of the fossils was sent to Yochelson, who identified them as *Salterella.* The location is the first well documented *Salterella* locality in Alabama; these fossils also are the first *Salterella* from Alabama for which detailed study is presented.

GEOLOGIC SETTING

The *Salterella* locality is near the leading edge of the Helena thrust sheet in the Appalachian fold-thrust belt in central Alabama (Figure 1). The Helena fault is a thrust fault of regional magnitude, and the Paleozoic succession in the hanging wall shows marked differences from that in the footwall (Butts, 1926; Szabo and others, 1988). The fault is interpreted as a major ramp from the basal decollement (Thomas, 1982, 1985). At Helena, the Helena fault juxtaposes Lower Cambrian rocks and Pennsylvanian rocks.

The locality of concern is within a cut of the CSX Transportation railway through the Lower Cambrian Rome Formation on the north side of Buck Creek in Helena, Shelby County, Alabama (NW1/4 NE1/4 sec. 15, T. 20 S., R. 3 W.). The cut does not expose the Helena fault, but its location can be inferred between outcrops of the Rome Formation in the cut and Pennsylvanian rocks just west of the Rome exposure. Within the cut rocks dip to the east-northeast at an average of 50 degrees. Small-scale folding and faulting are locally present within the exposure, but stratigraphic separation on these minor faults is small and there does not appear to be significant repetition of beds.

Approximately 105 meters of the upper part of the Rome Formation is exposed in the cut (Figure 2). The sequence can be subdivided into three parts (Osborne, in press). The lower 35 meters is characterized by grayish-red-purple and olive-gray shale and mudstone containing minor interbedded olive-gray to dark-gray finely crystalline dolomite. This lower part of the exposed Rome grades upward to a 33-meter-thick sequence of dark-greenish-gray and grayish-red-purple shale and mudstone containing abundant interbeds of dark-gray finely crystalline dolomite. Near the middle part of the section are three thin (0.5 to 1.0 meter) beds of limestone containing *Salterella* in great abundance (Figure 3). A few fragments of trilobite carapaces also occur.

Within the limestone, several layers occur in which the fossil content is about 50 percent of the total volume (Figure 3). In these layers there is a preferential orientation to most conchs, and some conchs are imbricated.

The upper part of the exposed Rome is 37 meters thick and includes two bluish-gray and brownish-gray, very fine- to fine-grained sandstone sequences separated by an interval dominated by olive-gray shale. The shale sequence locally contains abundant *Olenellus* sp., inarticulate brachiopods, and possible hyolithids. The uppermost sandstone at the top of the Rome is overlain by carbonate residuum, probably from the Middle to Upper(?) Cambrian Brierfield Dolomite (Osborne, in press).

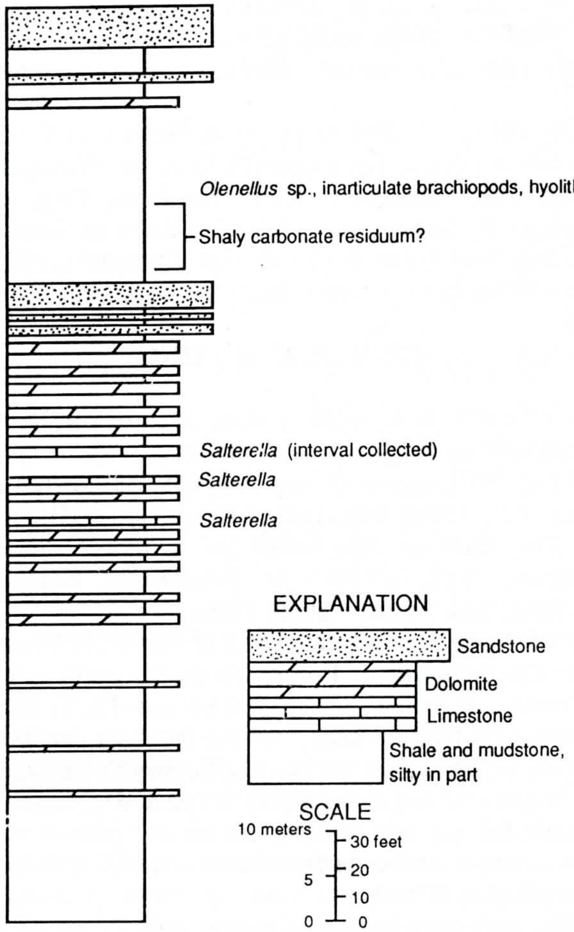

Figure 2. Generalized columnar section of the Rome Formation exposed at Helena, Shelby County, Alabama. Stratigraphic level of fossiliferous beds shown on right side of column.

STRATIGRAPHIC SIGNIFICANCE

For about a century, it has been recognized that a basal carbonate-clastic sequence within the Appalachians (Shady Dolomite, Rome Formation, and equivalent units) is of Early Cambrian age. However, by the description of specimens made from thin sections, this paper constitutes the final documentation that the same fossil occurs from Labrador to Alabama, ranging more than 3000 kilometers. In western North America, *Salterella* has a restricted stratigraphic range and serves as a good guide to rocks within the medial part of the *Bonnia-Olenellus* Zone (Fritz and Yochelson, 1988). Although the stratigraphic evidence in eastern North America is not as satisfactory, it is believed that the genus is similarly restricted stratigraphically.

For detailed work on the Shady Dolomite and its correlatives, such as the Tomstown Dolomite (Reinhart and Wall, 1975), occurrences of *Salterella* mark a potentially key zone for correlation. The occurrence of this fossil in limestone beds within the Rome Formation is equally important in potentially linking, by the

Figure 3. Side view of a specimen of *Salterella*-bearing limestone showing several bands of fossils within the rock. USNM 440899 (U. S. National Museum of Natural History specimen number).

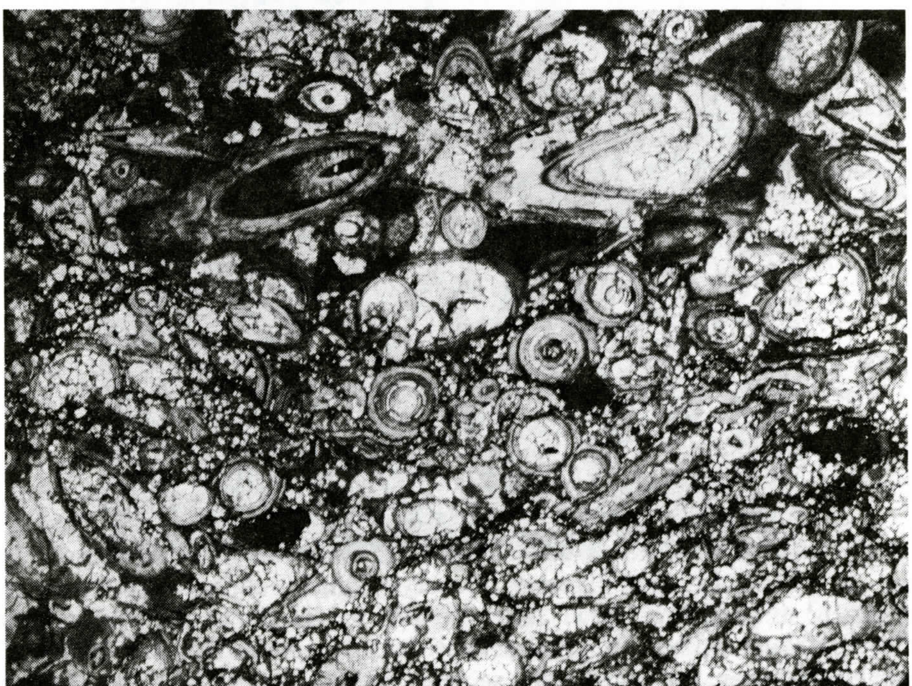

Figure 4. Thin section through a hand specimen of limestone showing mainly cross sections and oblique cross sections of *S. maccullochi*. Magnification X 10. USNM 440899.

means of the same fossil, facies within the Rome Formation and Shady Dolomite.

SYSTEMATIC PALEONTOLOGY

Phylum Agmata Yochelson, 1977
Family Salterellidae Walcott, 1886
Genus *Salterella* Billings, 1861
Salterella maccullochi Murchison, 1859

Description

The specimens from Alabama are typical for the species. In cross section (Figure 4) they show the characteristic "bullseye" pattern of the outer wall surrounding the inner deposits, and in turn either a large inner circle if the specimen is cut near the aperture, or a small circle if the central tube is cut. The inner deposits in many individuals show a slightly different color and texture, even under a hand lens, than the outer wall. Specimens cut obliquely show concentric ovals.

Transverse sections show the "V" shape of the exterior, without any rounding near the apex (Figure 5). The relationship of the inner laminae to the outer wall

Figure 5. Thin section showing a very slightly oblique specimen of *S. maccullochi* in transverse view. The apertural cavity shows the laminated wall; part of the central tube is represented by the elongate white area below. The distinction between the outer wall, secreted by the animal, and the inner wall, composed of collected particles cemented in place, shows as a difference in shade of gray. This is best seen on the upper right side where a minor fault displaces the apical area of a small broken specimen. Magnification X 25. USNM 440899.

is best observed in this view. The volume of the inner laminae and the distance that they extend away from the apex depend entirely on the size of the specimens. As with other examples, the smallest specimens contain no inner laminae, but because the amount of filling increased more rapidly than the specimens grew, the largest individuals are about two-thirds filled with laminae, excluding the hollow central tube. On most specimens the apertural rim is sharp, indicating little or no post mortem wear.

Discussion

A reillustration of *S. maccullochi* (Murchison) based on topotypes has been given by Yochelson (1983). As knowledge of *Salterella* and especially of the degree of its individual variation has increased, the number of species names has decreased. Different designations in the literature commonly refer to the same entity depending on when the work was published. Thus, *Salterella rugosa* Billings from Labrador, as refigured by Yochelson (1977), unfigured material from southern Pennsylvania (Yochelson, 1970), and *Salterella* new species from southwestern Virginia (Byrd, Weinberg, and Yochelson, 1973), all document occurrence of *S. maccullochi* along the length of the Appalachian Mountains. *S. maccullochi* always occurs immediately above, or a very short stratigraphic distance above, *S. conulata* Clark (see Yochelson, 1970 for redescription of that species) where that species is present in the section.

Yochelson (1983) has noted that J. W. Salter possibly was the person to describe this species; Billings (1861) assigned the authorship to Salter. For purposes of nomenclature, however, Murchison (1859) is the author.

ACKNOWLEDGEMENTS

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RELATIVELY UNALTERED WOOD OF *TAXODIUM DISTICHUM* FROM A PROBABLE EOCENE-AGED SINKHOLE NEAR CHATTANOOGA, TENNESSEE

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ABSTRACT

Relatively unaltered wood from *Taxodium distichum* was recovered from a sinkhole in the Knox Dolomite near Chattanooga, Tennessee. This wood is probably from the Eocene Epoch.

INTRODUCTION

Taxodium distichum (L.) Richard (bald cypress) primarily occurs on the coastal plain in the southeastern United States in low, wet areas (Elias, 1980). In fossil form it is a well-known Tertiary plant that is chiefly represented by foliage and cones; fossils of *Taxodium* wood have not been widely reported, although they have been described from the Eocene of the Porcupine and Great Valley groups (Penhallow, 1907).

During the expansion of a shopping center located at the intersection of US Highway 11 and Interstate 152 east of Chattanooga (Figure 1), the Stein Construction Company of Chattanooga, Tennessee, uncovered several fragments of wood from a clay deposit within the Knox Dolomite (Wilson, 1989). The woody material consisted of several fragments 50 by 75 mm. It was encased in a very stiff bluish-gray deposit of clay surrounded by typical cherty clays of the Knox group. The wood was encountered some 12 m beneath the surface in an ancient sinkhole. Bauxite deposits throughout the Valley and Ridge have been reported from ancient sinkholes by Adams and others (1926) and Bridge (1950). Sears (1957) reported the presence of lignite in late Cretaceous or early Tertiary erosion surfaces that formed over the Shady Dolomite in southwest Virginia. Depth of burial of the wood, together with the surrounding nature of the clay-like deposit that resembled many of the described Tertiary-aged bauxite deposits in the area (McIntosh, 1949), lead us to believe that this wood was part of an ancient sinkhole deposit formed on the Knox Dolomite during Eocene time.

The wood specimens were in such good condition that they were sent in for radiocarbon analysis to be certain that they were not recently placed at the site. A person could not ascertain by cutting and examining the wood that it was not just

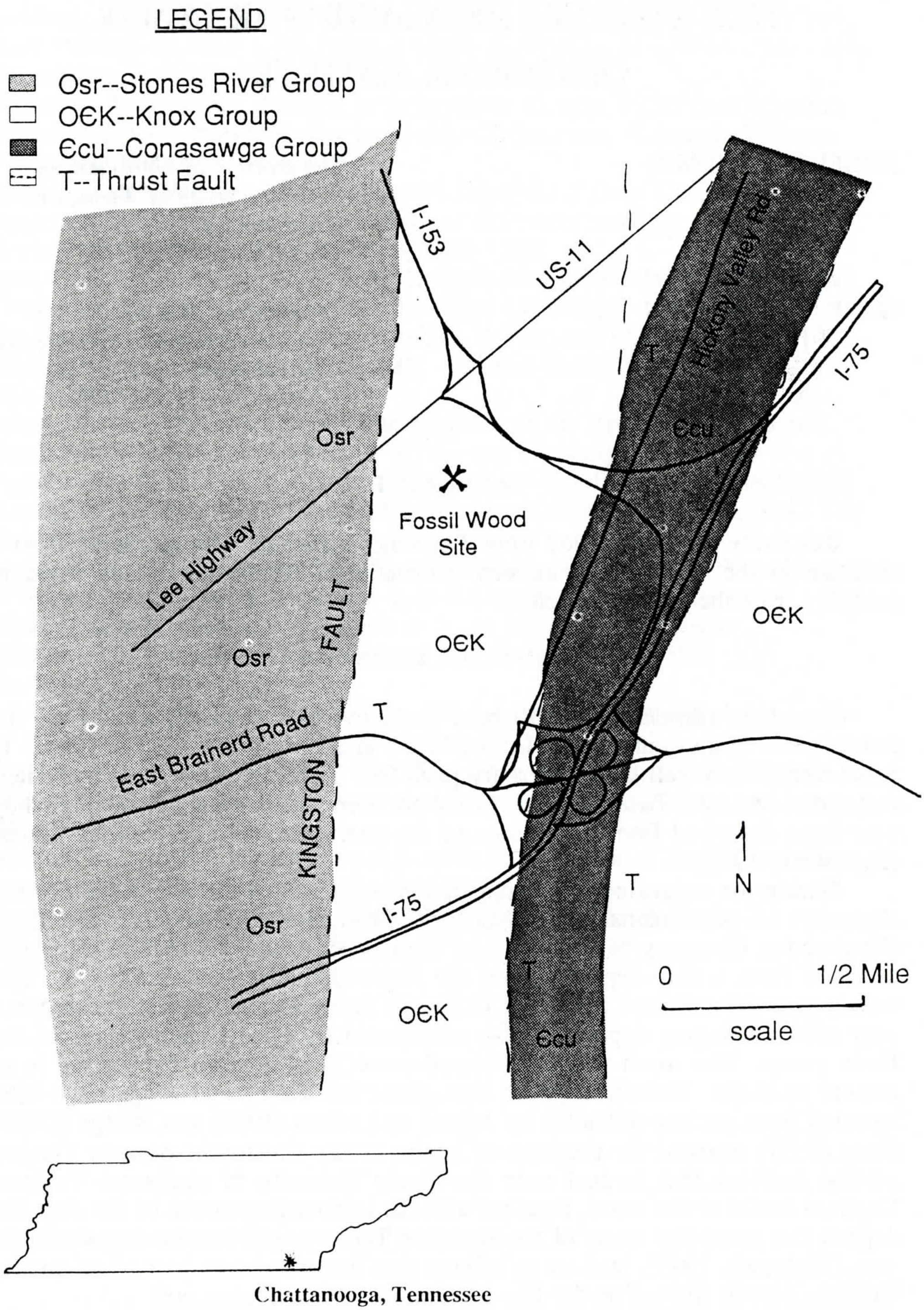

Figure 1. Location Map.

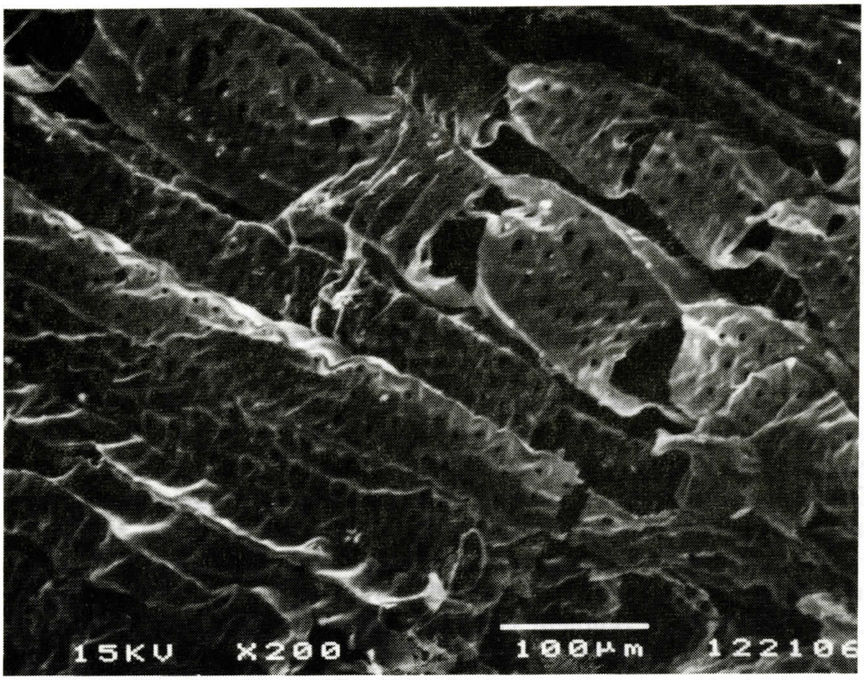

Figure 2. Scanning electron micrograph of internal view of tracheids from broken wood showing its excellent condition.

a few years old. That the wood was not recent was confirmed by Teledyne Isotopes of Westwood, New Jersey, which indicated an age of 40,000 plus years for a sample. This specimen of wood, which we identified as bald cypress, *Taxodium distichum*, had fallen into a sinkhole from the eroding surface of the Knox Dolomite in early Tertiary.

The peculiar part of the find is that the wood is in very good condition. Jeffrey (1904) reports finding wood of similar condition from the Miocene. His find is described as *Sequoia penhallowii* Jeffrey and was noted as being closely related to *Sequoia gigantea* Lindl., now in the genus *Sequoidendron.* Like *Taxodium*, *Sequoia* and *Sequoidendron* are in the Taxodiaceae.

METHODS

The wood was examined by both light microscopy and scanning electron microscopy. Material was macerated using Jeffrey's method (Johansen, 1940), but only kept in macerating fluid for 24 hours. Some material was left unstained and some was stained with safranin. Material to be viewed with light microscopy was mounted with diaphane.

Material to be viewed by scanning electron microscope was mounted on metal stubs with Scotch double-sided tape. The material was then coated with gold using a Hummer VI Sputtering System. Stubs were coated for two minutes at 10 millivolts. A JEOL JSM-T220A scanning electron microscope was used for observation of both macerated and broken material.

RESULTS

The wood is from *Taxodium distichum* as determined using a key in Brown, Panshin, and Forsaith (1949). Key features are a lack of vessels, lack of resin canals, lack of spiral thickening in tracheids, and presence of pits in the ray crossings of spring wood that are oval-orbicular and 7.5 microns in their longest direction. In addition to use of the key, the material was compared with prepared microslides of *Taxodium distichum* (L.) Richard, *Sequoia sempervirens* (D. Don) Endl., and *Sequoiadendron gigantea* (Lindl.) Buchholz purchased from Ripon Microslides in Wisconsin. A scanning electron micrograph is shown in Figure 2.

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